# Surface Heat Flux Analysis of Onondaga Lake

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S.Haran

# Abstract

Accurate prediction of the water surface heat flux in a lake in many cases is an essential prerequisite for developing a comprehensive water quality model. The five essential components of the surface heat flux, which are solar radiation, longwave atmospheric radiation, longwave back radiation, evaporative and conduction are systematically evaluated using the predictive formulae presented by Orlob et al. (1983), Livingstone and Imboden (1989), Imberger and Patterson (1981), Henderson-Sellers (1986), and Environmental Laboratory (1986) for the study period 1985-1991 for Onondaga Lake. Direct measurements of incident solar radiation were available for portions of this time period. Five alternative solar radiation formulae namely Environmental Laboratory (1982), Environmental Laboratory (1986), Krambeck (1982), Brock (1981), and Henderson-Sellers (1986) are investigated. Each of them are of different complexcity and empiricity. A simple statistical test of root mean square (r. m. s. ) is performed with the computed values and the measured values during the period of valid measurements to arrive at a statistical conclusion. Environmental Laboratory (1982) performed adequately with a mean r. m. s. value of 33.8 $Kcal/m^2/hr$. The averaged lake temperature obtained on days of temperature profile measurements and the subsequent evaluation of the net surface heat flux between those successive days is used as the basis for comparing the net surface heat fluxes as calculated from the five recommendations earlier mentioned. The root mean square (r. m. s. ) error was computed to arrive at a statistical conclusion. Livingstone and Imboden (1989) performed adequately with a mean r. m. s. value of 82 $Kcal/m^2/hr$.

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# Chapter 1 INTRODUCTION

In most water quality models of surface impoundments, the surface heat flux plays an important role in initiating the complex mixing mechanisms (Henderson-Sellers, 1986). Usually the heat flux between the sediment-water interface is significantly less and is often ignored in most studies (Livingstone and Imboden, 1989). The heat flux associated with the inflows and the outflows are also often neglected especially if the residence time associated with them are small. The inflows and outflows enter and leave the surface waters of the lake.

Figure 1 shows the surface heat fluxes associated with the heat budget of any water body. The internal thermal mixing processes are partly the direct consequence of this water-air energy exchange. The cooling of the surface layer relative to the lake body will enhance the mixing processes as a result unstable density variations. Any warming will bring about a more stable water column that is resistent to vertical mixing. The radiative components of the fluxes which includes the incident solar radiation, longwave atmospheric radiation, and longwave back radiation are usually the larger components of the heat budget model and an increase in these preceeds any increase in the non-radiative component (Livingstone and Imboden, 1989). The non-radiative components of the heat budget are the evaporative and conductive losses. The energy flux associated with precipitation is negligible (Henderson-Sellers, 1986). While all the surface heat fluxes are an air-water interaction, the incident solar radiation is able to penetrate the depths and heat the layers below the upper $1-2$ mm of water body. The degree to which this penetration occurs is dependent on the extinction coefficient of the upper water columns.

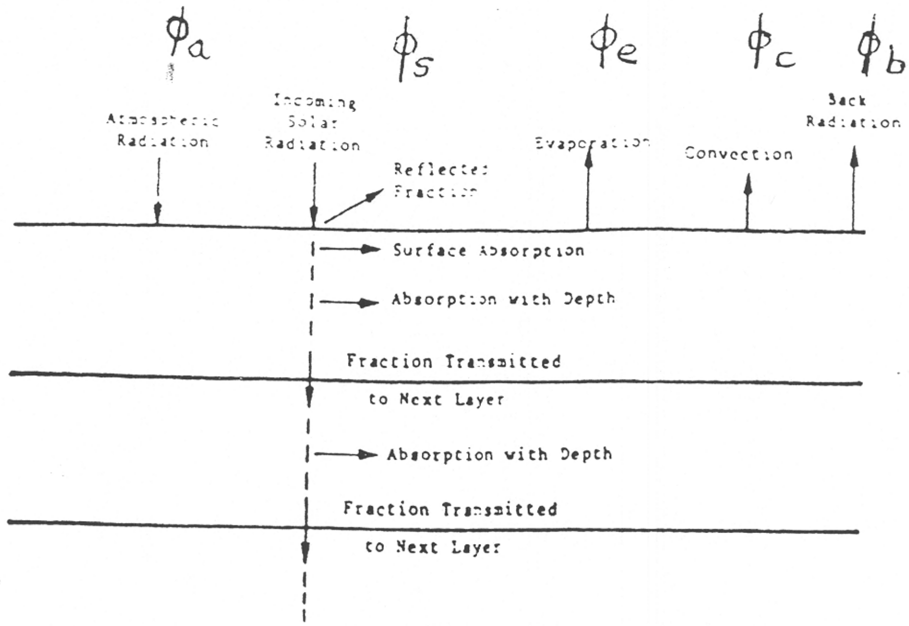


Figure 1: Schematic diagram of source and sink terms in the heat budget model (Source: Hondzo and Stefan, 1991)

This extinction coefficient is a cumulative index of all the dissolved/suspended organic and inorganic particulates in the water. This demonstrates the coupled nature of the heat and mass conservation equations in water quality modeling studies.

Every lake has a unique heat budget pattern that is dependent on its geography, climate, sources and sinks of heating, lake morphometry and the penetrative shortwave energy component across the the depths of the water column (Wiegand et al. , 1982). As Jorgensen et al. (1982) suggests, many of the water quality modeling formulations do not always use the best available formulae, but rather make a random choice without any prior study or justification. It is hoped that that this study would answer those concerns and lay the framework for future studies of Onondaga Lake.

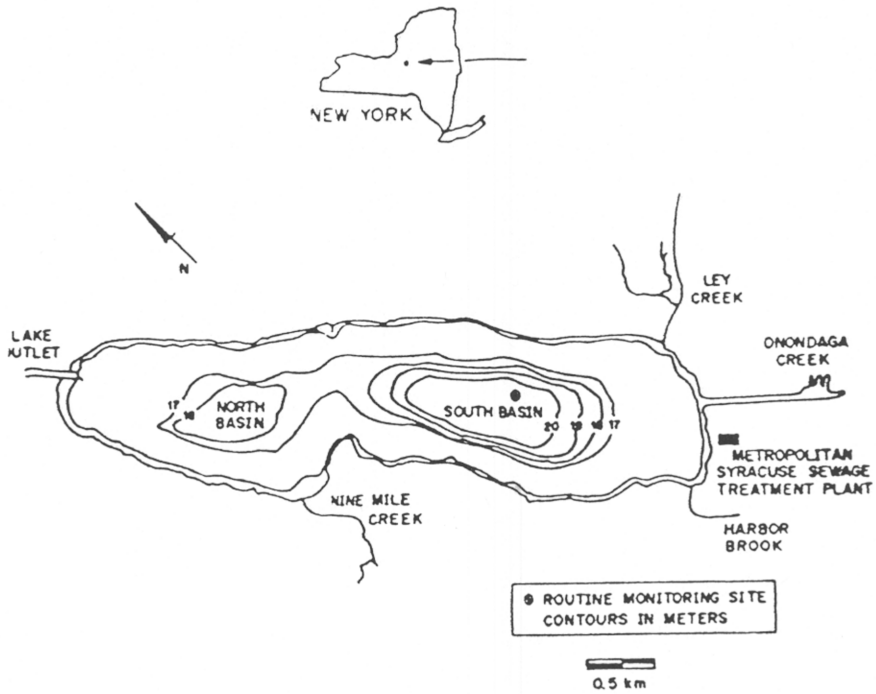


Figure 2: Location of Onondaga Lake in New York state (Source: Canale and Effler, 1989)

## 1.1 Onondaga Lake

Onondaga Lake is a shallow dimictic lake located adjacent to the city of Syracuse. It has the following geographic coordinates: latitude 42.9 deg N; longitude 76 deg W; and elevation 410 feet. Stable stratification occurs every summer. The maximum depth of the lake is about 19 m. From the definition given in Fox et al. (1979), it could be given the temperate lake classification as it is located at a latitude greater than 40 deg N.

A number of studies of the lake have been conducted (Devan and Effler, 1983; Canale and Effler, 1989). Due to the pollution from such sources as chloro-alkali manufacturer and the Metro waste treatment plant, episodes of anoxic conditions prevail during lake stratification. The bathymetric and geographical map is given in Figure 2. The morphometric data of Onondaga lake is given in Table 1.

Table 1: Morphology of Onondaga Lake (Source: Devan and Effler, 1983)

| Discriptor (1) | Value (2) |
|---|---|
| Drainage basin area | 600 $km^2$ (234 sq mile) |
| Lake surface area | 11.7 $km^2$ (4.6 sq mile) |
| Lake volume | $1.41 \times 10^8$ $m^3$ ($4.82 \times 10^9$ cu ft) |
| Mean depth | 12.0 m (39 ft) |
| Maximum depth | 20.5 m (66.6 ft) |
| Shore line length | 12.9 km (11.2 mile) |

## 1.2 Objective of the research

The first objective is to investigate the accuracy of alternative solar radiation formulae. This is analysed by comparing predictions to direct measurements made at Syracuse, N. Y. The second objective is to investigate the accuracy of alternative formulae for the prediction of the remaining components of the surface heat flux. This study is carried out by comparing predictions of net surface heat flux to the measured values as derived from water temperature profile measurements. The duration of the study for Onondaga lake is limited to spring-autumn periods of 1985–1991. The meteorological data used in the study was obtained from records kept by NOAA at Syracuse's Hancock airport. The incident solar radiation was measured by the Upstate Freshwater Institute.

# Chapter 2 Literature Review of Similar Work done on Lakes

## 2.1 Review of Surface Heat Flux Studies and Solar Radiation

Solar radiation is the driving force behind all the components of the surface heat flux. It diffuses and scatters as it reaches the surface of the earth, as determined by the prevailing atmospheric and meteorological conditions. The longwave atmospheric radiation is related to the cloudcover due to the emission-effect of the clouds. The incoming radiation heats up the surface water temperature depending on the mixing conditions of the lake. This in turn determines the evaporative, conductive and back-radiation capacity of the surface waters as dictated by the prevailing meteorological and the sunsequent mixing conditions in the lake.

*Surface Heat Flux Studies*

In Dutton and Bryson (1962), the heat flux terms of lake Mendota was studied. This was one of the earliest such work to sytematically quantify and qualify the heat content of a lake in order to understand the hydrothermal mixing processes prevalent in a lake.

In Myrup et al. (1979), the average monthly energy and water budgets are coalesced to support and correct the respective observations for Lake Tahoe, California-Nevada. The annual energy budget was found to be dominated by the net radiation and evaporation terms with 93% of the radiation input going to evaporate the water.

In Wiegand et al. (1982), the interaction between the surface and the advective component of the heat content of Lake Kootenay, British Columbia, Canada is investigated. It was found that in spite of the fact that the lake has large through-flow, the heat budget was more in character with a lake that is only slightly influenced by rivers.

In Henderson-Sellers (1986), a systematic review is made for the five components of the surface heat flux and successfully applied to a lake in South Africa and United Kingdom. Based on empirical and theoretical evidences, recommendations are made for each of the five components. The review analyses all possible equations used in modeling studies and comes up with possible justification for each selection.

In Livingstone and Imboden (1989), the annual heat balance and the subsequent equilibrium temperature of Lake Aegeri, Switzerland was investigated. Only the direct (i. e. latent heat of fusion) influence of the icecover partially or fully covering the lake during winter was considered. The through-flow term was incorporated in to the study and was found to be negligible to the heat content of the lake.

It is worth noting that much of the earlier heat-budget work led to better evaporation formulations and contributed to the rapid evolution of water quality modeling of lakes and reservoirs.

*Solar Radiation*

The sun is treated as a black body with the relative distance between the earth and the sun which varies as a function Julian day, taken care of by the inverse square law. The earth's rotation around its axis and its revolution around the sun produces a daily cycle and annual cycle respectively in the received solar radiation at the top of the atmosphere. However the incident solar radiation at any geographic location on earth is a function of Julian day, latitude, altitude, vegetation cover, cloud cover, dust particles emitted by both anthropogenic and natural sources, and atmospheric conditions, and the three latter parameters are the weak links in the empirical formulations of most models as they are essentially random properties or difficult to measure. Most models explicitly or implicitly incorporate these physical concepts as well as the randomness of some of the parameters into the model formulation. Angström (1922) suggested a relationship based on the number of sunspots during that year, but these are rarely incorporated in the ecological modeling field. Figures 3 and 4 illustrate the latitudinal and seasonal dependance of solar radiation at the top and surface of the earth. Figure 5 illustrates the received solar radiation during the June soltice as a function of latitude. Figure 6 illustrates altitudinal dependence of solar radiation as observed in the European Alps. Figure 7 illustrates the transmission properties of different cloud thichnesses. Figure 8 illustrates the diurnal variations of radiant energy and temperature in the middle and low latitudes. The most common instrument for measuring solar radiation is the pyranometer. In essence, it is a thermocouple attached to a black surface that is adequately calibrated to measure the incident solar radiation.

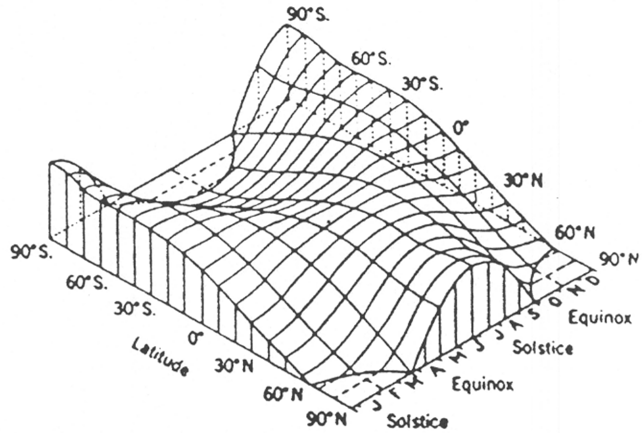


Figure 3: The variation of solar radiation with latitude and season assuming no atmosphere Source: Barry and Chorley, 1987

## 2.2 Prediction of Solar Radiation

### 2. 2. 1 *Brock (1981) recommendation*

Brock (1981) suggested the modified Angtröm-Prescott approach for treating the solar radiation as a computed variable in the field of ecological modeling. This formulation is probably the most empirical and requires a previous study at or near the locality. However, it has the advantage of simplicity and ease of use, and requires no input of current meteorological data. The climatic constants utilized in the approach have been determined for a large number of sites around the world by deJong (1973). The solar radiation formulation is given by:

$$\phi_s = \phi_\infty(a + b\frac{n}{D}) \tag{1}$$

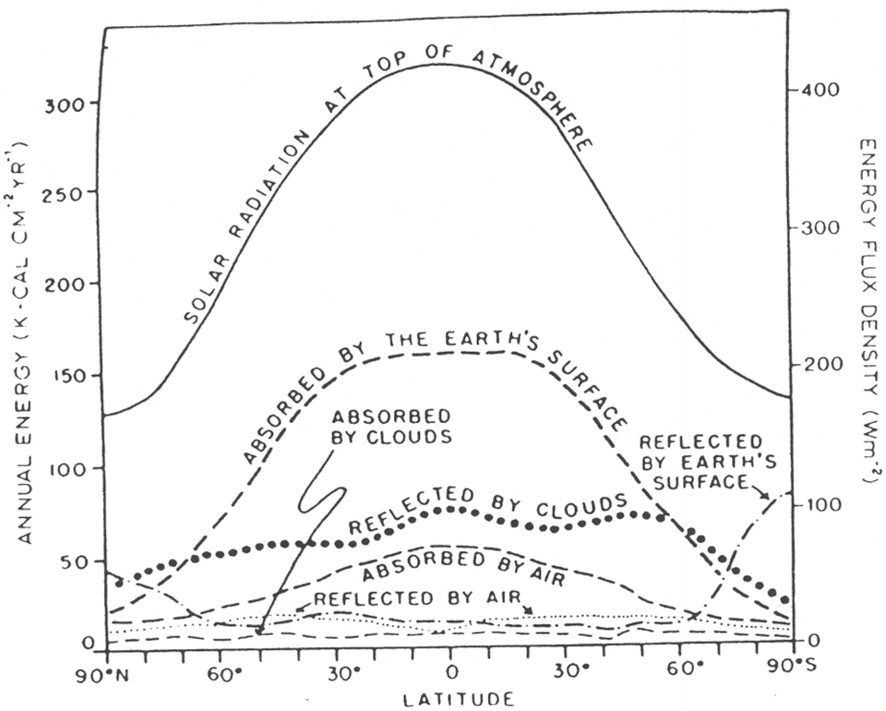


Figure 4: The average annual latitudinal disposition of solar radiation (Source: Barry and Chorley, 1987)

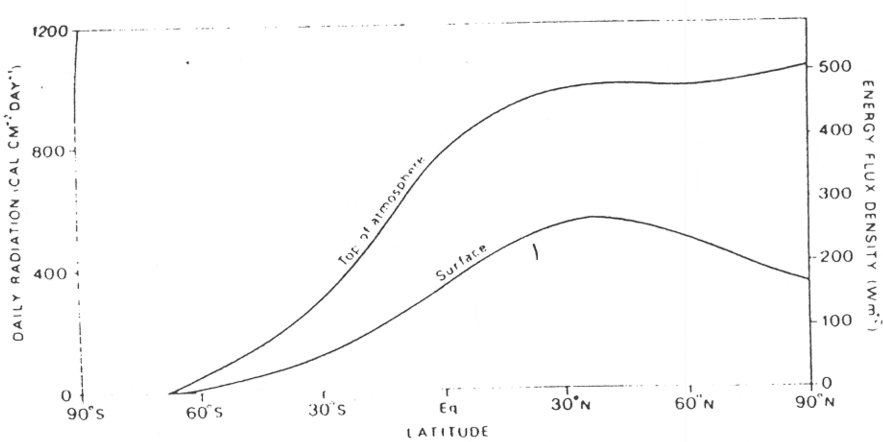


Figure 5: The average receipt of solar radiation with latitude at the top of the atmosphere and at the surface during the June soltice(Source: Barry and Chorley, 1987)

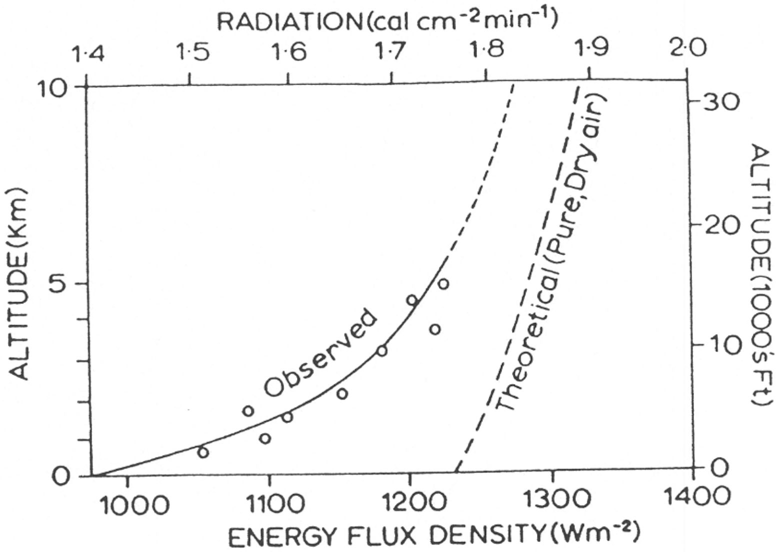


Figure 6: Direct solar radiation as a function of altitude observed in the European Alps (Source: Barry and Chorley, 1987)

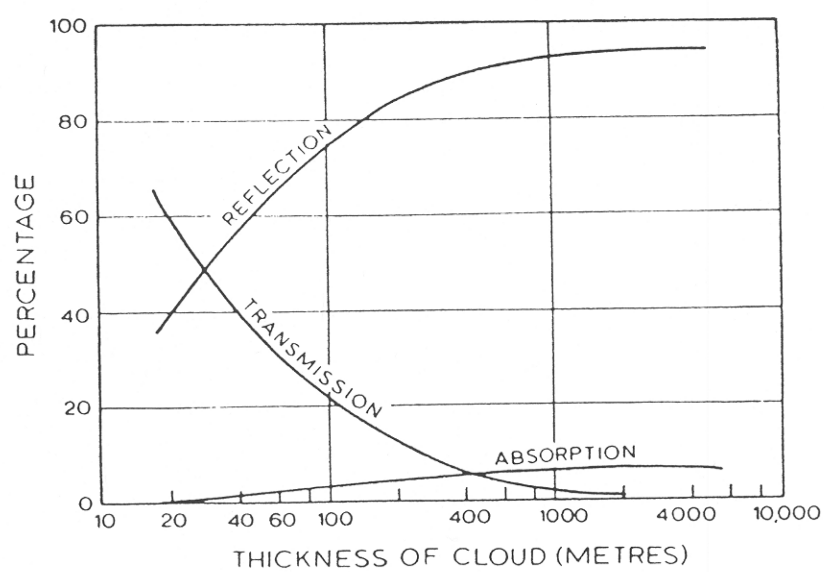


Figure 7: Percentage of reflection, absorption, and transmission of solar radiation by cloud layers of different thickness (Source: Barry and Chorley, 1987)

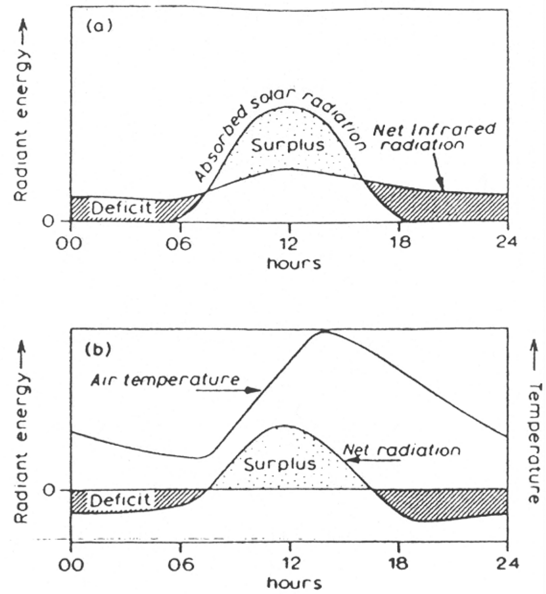


Figure 8: Curves showing diurnal variations of temperature and radiant energy A. Diurnal variations in absorped solar radiation and infra-red radiation in the middle and low latitudes B. Diurnal variations in net radiation and air temperature in the middle and low latitudes. (Source: Barry and Choley, 1987)

where $\phi_\infty$ is solar radiation at the top of the atmosphere, ($a$ & $b$) are the climatic empirical constants,

$\frac{n}{D}$ is duration of sunshine, $\frac{n}{D} = 1 - C$ (as suggested by Henderson-Sellers (1986)), and $C$ is cloud cover.

The solar radiation at the top of the atmosphere is given by:

$$\phi_\infty = \left[\frac{24}{\pi}\right]\left[\frac{I_0}{R1}\right]\left[\left(\frac{W1\pi}{180}\right)\sin(L)\sin(\delta) + \sin(W1)\cos(L)(\cos(\delta))\right] \qquad (2)$$

where $I_0$=1353 $Wm^{-2}$ is the solar constant, $W1$ is the sunset angle, $R1$ is the radius vector, $\delta$ is the declination angle and $L$ is the latitude.

The sunset angle, radius vector, and the angle of declination are given by the equations

3, 4 and 5.

$$W1 = \arccos\left[-\left[\tan(L)\tan(\delta)\right]\right] \tag{3}$$

where $W1$ is sunset angle.

$$R1 = \frac{1}{\left(1 + 0.33\cos\left(\frac{360JDAY}{365}\right)\right)^{0.5}} \tag{4}$$

where $R1$ is radius vector, $L$ is latitude, $\delta$ is declination angle.

$$\delta = 23.45\sin\left[360\left(\frac{284 + JDAY}{365}\right)\right] \tag{5}$$

The listing of empirical climatic constants for different locations is given in deJong (1973). The constant $a$ is assoociated with the latitude and the constant $b$ is associated with the atmospheric turbidity (Henderson-Sellers, 1986).

2. 2. 2 *Krambeck (1982) recommendation*

Krambeck (1982) suggests a formulation that is equally simple but rarely been tried outside the intended site of Plön, West Germany, for which it was designed. It utilizes only the latitude and the cloudcover data, which can be obtained using satellite data if it the need arises. It is perhaps the most deterministic in its theoretical formulation.

The actual earth-sun distance, $R$, is given by:

$$R = R_0\frac{1 - e^2}{1 - e\cos(\varphi - \varphi_a)} \tag{6}$$

where $R_0$ is $1.495 \times 10^{13}$ cm (the average earth-sun distance), $e$ is the numerical eccentricity of the earth's orbit (0.017), $\varphi_a$ is the angle of aphelion (181.5 deg ), and $\varphi$ is the angle of earth's revolution around the sun.

The angle of sun's elevation is given by:

$$\sin(\gamma) = \sin(b)\sin(\delta) - \cos(b)\cos(\delta)\cos(15t) \tag{7}$$

where $\gamma$ is the angle of sun's elevation, $b$ is the latitude, $\delta$ is the angle of declination and $t$ is the time of simulation.

The angle of declination at any point of earth's revolution around the sun is given by:

$$\sin(\delta) = \sin(\delta_0)\cos\left(\varphi - \varphi_a\right) \tag{8}$$

where $\delta = 23.45$ deg.

The relative path of light, $L(\gamma)$, as utilized in the formulation is given by:

$$L(\gamma) = a \int_0^{\infty} \frac{exp(a\ D)\ \ (R_e + D)}{\left(R_e{}^2 \sin^2(\gamma) + 2R_e D + D^2\right)^{0.5}} dD \tag{9}$$

where $a$ is the atmospheric constant (0.125122 $Km^{-1}$), $L(\gamma)$ is the relative path of light, $D$ is the thickness of atmosphere, and $R_e$ is the radius of earth ($6.37 \times 10^8$ cm).

The incident solar radiation is given by the equation given below:

$$\phi_s = I_0 \left(\frac{R_0}{R}\right)^2 exp(-C\ \ L(\gamma))\ \sin(\gamma) \tag{10}$$

2. 2. 3 *Henderson-Sellers (1986) recommendation*

Henderson-Sellers (1986) suggests using the Tucker (1982) recommendation for the solar radiation formula. He also suggests using the modified Angström-Prescott approach if the local meteorological observations are not available. The solar radiation is given by:

$$\phi_s = (\phi_{sd} + \phi_{ss})[1 - (1 - k')C] \tag{11}$$

where $\phi_{sd}$ is the direct component of solar radiation, $\phi_{ss}$ is the scattered component of solar radiation, and $k'$ is a coefficient of latitude (see Henderson-Sellers, 1986).

The direct component of the solar radiation is given by:

$$\phi_{sd} = I_0 \sin(\alpha) F_s^{\theta_{am}} \tag{12}$$

where $\alpha$ is the solar altitude, $F_s$ is the atmospheric term, and $\theta_{am}$ is the optical air mass.

The last three terms are substituted with the similar terms found in the Environmental Laboratory method (see sections 2. 2. 4 and 2. 2. 5).

The scattered component of solar radiation is given by equation shown below:

$$\phi_{ss} = 0.38(I_0 - \phi_{sd}) \sin(\alpha) \tag{13}$$

The *albedo* is given by:

$$albedo = \frac{a_0}{a_0 + \sin(\alpha)} \tag{14}$$

where $a_0$ is a variable.

This variable is computed using the equation shown below:

$$a_0 = 0.02 + 0.01\Big[0.5 - C\Big]\left[1 - \sin\Big(\frac{\pi(JDAY - 81)}{183}\Big)\right] \tag{15}$$

2. 2. 4 *Environmental Laboratory (1986) recommendation*

Environmental Laboratory (1986) provides a complicated set of procedures for evaluating the solar radiation.

The solar radiation is given by:

$$\phi_s = F_r\ F_c\ F_s\ \frac{I_0}{R^2} \sin(\alpha) \tag{16}$$

where $F_r$ is the reflection factor, $F_c$ is the cloudiness factor, $F_s$ is atmospheric transmission term, $I_0$ is the solar constant ($\approx 0.33 Kcal/m^2/sec$), $R$ is the relative earth-sun distance, and $\alpha$ is the sun angle (rads).

The relative earth-sun distance (distinct from the radius vector defined in section 2. 2. 1) is given by the equation shown below:

$$R = 1 + 0.17\cos\left[\frac{2\pi}{365}(186 - JDAY)\right] \tag{17}$$

The atmospheric transmission term is given by:

$$F_s = \frac{a'' + 0.5(1 - a' - d)}{1 - 0.5k(1 - a' + d)} \tag{18}$$

where $a''$ is the mean atmospheric coefficient after scattering and absorption, $a'$ is the atmospheric coefficient, $d$ is the (adjustable) dust attenuation coefficient, and $k$ is the albedo.

The mean atmospheric coefficient after scattering and absorption is given by:

$$a'' = exp\left[-\Big[0.465{+}0.0408\Big(0.00614exp\Big(0.0489T_d\Big)\Big)\Big]\Big[0.179{+}0.421exp\Big(-0.721\theta_{am}\Big)\Big]\theta_{am}\right] \tag{19}$$

where $T_d$ is the dew-point temperature [deg $F$], and $\theta_{am}$ is the optical air mass.

The optical air mass is described by Markofsky and Harleman (1971) as the ratio of the path length of the sun's rays through the atmosphere to the path when the sun is directly overhead.

The optical air mass is given by:

$$\theta_{am} = \frac{exp\left(\frac{-ALT}{2532}\right)}{\sin(\alpha) + 0.15\left(\frac{180\alpha}{\pi}\right)^{-1.253}} \tag{20}$$

where $ALT$ is the elevation of the site.

The atmospheric coefficient is given by the equation shown below:

$$a' = exp\left[-\Big[0.465 + 0.0408\big(0.00614exp\big(0.0489T_d\big)\big)\Big]\Big[0.129 + 0.171exp\big(-0.88\theta_{am}\big)\Big]\theta_{am}\right] \tag{21}$$

The cloudiness factor is given by the equation shown below:

$$F_c = 1 - 0.65C^2 \tag{22}$$

The reflection factor is given by the equation shown below:

$$F_r = 1 - albedo \tag{23}$$

The albedo function is given by:

$$albedo = A(57.3\alpha)^B \tag{24}$$

where $A$ and $B$ varies as a function of cloud cover as shown below:

$$A = 1.18 \text{ and } B = -0.77 \text{ if } C < 0.05$$

$$A = 2.20 \text{ and } B = -0.97 \text{ if } 0.05 \leq C < 0.5$$

$$A = 0.95 \text{ and } B = -0.75 \text{ if } 0.5 \leq C < 0.95$$

$$A = 0.35 \text{ and } B = -0.45 \text{ if } 0.95 < C$$

The solar angle in rads is given by:

$$\sin(\alpha) = \sin\phi\sin\delta + \cos\phi\cos\delta\cos\omega \qquad \alpha \geq 0.01 \tag{25}$$

where $\phi$ is the latitude (rads), $\delta$ is the solar declination (rads), $\omega$ is the solar hour angle (rads).

The declination angle in rads is given by the equation shown below:

$$\delta = 0.4092\cos\left[\frac{2\pi}{365}\left(172 - JDAY\right)\right] \tag{26}$$

The solar hour angle is given:

$$\omega = \frac{\pi}{12}\left(t - t_L - 12\right) \tag{27}$$

where $t$ is the simulation hour, and $t_L$ is a fraction of a 15 deg increment by which the local meridian is west of the standard meridian for the time zone $\approx \frac{1.31}{15}$ for Syracuse.

The sun-set hour ($t_{ss}$) and the sun-rise hour ($t_{su}$) are given by the equations 28 and 29 respectively:

$$t_{ss} = \frac{12}{\pi}\arccos\left(\frac{-\sin\phi\sin\delta}{\cos\phi\cos\delta}\right) + t_L + 12 \tag{28}$$

$$t_{su} = -t_{ss} + 2t_L + 24 \tag{29}$$

2. 2. 5 *Environmental Laboratory (1982) recommendation*

An earlier Environmental Laboratory (1982) method has the same formulation as the Environmental Laboratory (1986) method described above except in the configuration of the atmospheric transmission term $F_s$. The atmospheric transmission term, $F_s$, in terms of the (adjustable) turbidity factor as shown below:

$$F_s = exp\left[\frac{T(0.128 - 0.054\log(\sin\alpha))}{\sin\alpha}\right] \tag{30}$$

where $T$ is the turbidity factor.

## 2. 3 Prediction of other Components of Surface Heat Flux

For purposes of this study, all incoming radiations into the lake surface is considered positive while all outgoing radiation and losses are considered negative. Five sets of recommendations are utilized for the surface heat flux analysis:

1. Livingstone and Imboden (1989)
2. Imberger and Patterson (1981)
3. Orlob et al. (1983)
4. Environmental Laboratory (1986)
5. Henderson-Sellers (1986)

These were selected based on the most recent article/publication and the portability and adaptability of the prescribed variables to the available lake and meteorological measurements. It would be worth mentioning that the earlier pioneering work done at MIT and other places invariably led to the above mentioned articles using refinements to the original work.

### 2. 3. 1 *Solar radiation ($\phi_s$)*

Most recommendations prescribe the use of measured quantities of solar radiaion for diagnostic water quality model investigations. However as presented in the previous section, a substitute formula would be necessary when measurements are incomplete or non-existent. For prognostic modeling studies, the trend as evaluated from historical records or an alternative formula matching the past trend or predicting the anticipated future trend can be used.

### 2. 3. 2 *Back radiation ($\phi_b$)*

The radiation emitted by the surface water is considered to follow the black body radiation. There is almost a universal agreement as to the nature of the back radiation formula as given by the equation below:

$$\phi_b = -\epsilon\ \sigma T^4 \tag{31}$$

where $\sigma$ is the Stefan-Boltzmann constant ($2.0411 \times 10^{-7} KJm^{-2}hr^{-1}K^{-4}$), $T$ is the absolute temperature of lake surface $[K]$, and $\epsilon$ is the emissivity coefficient.

All except Orlob et al. (1983) and Imberger and Patterson (1981) recommend a value of 0.97 for the emissivity coefficient with the latter two recommending a value of 0.96.

### 2. 3. 3 *Atmospheric radiation ($\phi_a$)*

Atmospheric radiation from the sky is treated as a black body radiation. The basic differences in the respective formulations are in the emission coefficient ($E_L$) of the sky. There exists a considerable number of formulations for computing the atmospheric radiation flux. The earlier versions were derived empirically through in-direct studies of the atmospheric heat flux and measurements while the recent formulations basically added refinements to the former by adapting to a particular set of circumstances and locations.

Livingstone and Imboden (1989) recommends the following equation:

$$\phi_a = 0.97 E_L \sigma {T_a}^4 \tag{32}$$

where $\phi_a$ is the net atmospheric radiation, $T_a$ is the absolute temperature of air $[K]$, and $E_L$ is the emission coefficient, which is given by equation 33.

$$E_L = 1.09(1 + 0.17C^2)1.24(\frac{e_a}{T_a})^{\frac{1}{7}} \tag{33}$$

[Note: The emission coefficient is explained in terms of $T_a$ and $e_a$] The variable $e_a$ in equation 33 is the vapour pressure at $T_a$ in $[hPa]$. [Note: $hPa$ is numerically equivalent to $mbar$]

Imberger and Patterson (1981) recommend the following equations for the computation of atmospheric longwave radiation:

$$\phi_a = (1 + 0.17C^2)\phi_{La} \tag{34}$$

$$\phi_{La} = 0.937\sigma {T_2}^4(1 - R_a) \tag{35}$$

where $T_2$ is the temperature of air at $2m$ height$[K]$, and $R_a$ is the water surface reflectivity $\approx 0.03$. [Note: The emission coefficient is taken to be a constant value of 0.937]

Orlob et al. (1983) recommends the following equation:

$$\phi_a = C_{at}\sigma {T_2}^6(1 + 0.17C^2)(1 - R_a) \tag{36}$$

where $C_{at}$ is the Swimbank's coefficient ($0.938\times\ 10^{-5}$).

Environmental Laboratory (1986) recommends the following equation:

$$\phi_a = 1.23 \times 10^{-16}(T_a + 273)^6(1 + 0.17C^2) \tag{37}$$

where $\phi_a$ is the net atmospheric radiation $[Kcal\ m^{-2}sec^{-1}]$, and $T_a$ is the dry bulb air temperature [deg $C$]. [Note: The emission coefficient is explained in terms of the Swinbank's coefficient]

Henderson-Sellers (1986) recommends the following set of equations 38-41 in the computation of the atmospheric radiation. The net atmospheric radiation is given by

$$\phi_a = 0.97\phi_{ai} \tag{38}$$

where $\phi_{ai}$ is the incident atmospheric radiation.

The incident atmospheric radiation is given by

$$\phi_{ai} = \epsilon_a\sigma {T_a}^4 \tag{39}$$

where $\epsilon_a$ is the atmospheric emission coefficient.

The atmospheric emission coefficient $\epsilon_a$ is given by equation 40 or 41 depending on the prevailing cloudcover:

$$\epsilon_a = 0.84 - \frac{n}{D}(0.1 - 9.973 \times 10^{-6} e_a) + 3.491 \times 10^{-5} e_a \quad \text{for } \frac{n}{D} \geq 0.4 \tag{40}$$

$$\epsilon_a = 0.87 - \frac{n}{D}(0.175 - 29.92 \times 10^{-6} e_a) + 2.693 \times 10^{-5} e_a \quad \text{for } \frac{n}{D} \leq 0.4 \tag{41}$$

where $\sigma$ is the Stefan-Boltzmann constant ($5.6697 \times 10^{-8} W m^{-2} K^{-4}$), $\frac{n}{D}$ is the duration of sunshine ($= 1 - C$). [Note: The emission coefficient is explained in terms of $\frac{n}{D}$]

2. 3. 4 *Evaporative losses ($\phi_e$)*

There are a considerable number of formulae available in the literature for the computation of evaporative heat losses, each with its own historical evolution. Evaporative heat losses are greater in magnitude than the conductive heat losses and the latter is expressed as a function of the former via the Bowen's ratio. Theorectically evaporative losses are treated as negative, but there are occasions when they will be positive. This occurs as a result of specific meteorological and stability conditions when condensation cccurs. These are however small in numerical value and it is often left to the discretion of the modeller to condition the simulation to return zero values under such circumstances. Free evaporation accounts for the natural evaporative capacity of the water surface unaided by the removal effect of the winds. Winds accelerate the removal of water vapour from the water surface and this is generally termed the forced (wind-induced) convectional evaporation. Most formulae incorporate both the free (i. e. when $T_s > T_a$ under calm conditions) and forced (wind induced) convectional effects into the formulation. Evaporative heat fluxes are generally formulated as a function of the vapour deficit between the air and water temperatures and the bulk transfer coefficient defining the turbulent eddy coefficient under the prevailing atmospheric conditions. The basic differnces in the different formulations are in defining the latter in terms of the wind speed.

Livingstone and Imboden (1989) recommend the following equation for the evaporative heat losses:

$$\phi_e = -f(e_s - e_a) \tag{42}$$

where $f$ is the wind function, and $e_s$ is the saturated vapour pressure at the lake surface temperature $[hPa]$.

The wind function is given by

$$f = 4.8 + 1.98u_{10} + 0.28(T_s - T_a) \tag{43}$$

where $u_{10}$ is the wind speed at $10m$ above water surface $[ms^{-1}]$.

Orlob et al. (1983) recommends the following general relationship for the computation of evaporative heat losses via the evaporation rate $(E)$ given in equation 45:

$$\phi_e = -\rho_w LE \tag{44}$$

where $\rho_w$ is the density of water 998.2 $Kg\,m^{-3}$ at $20\deg C$, $E$ is the evaporation rate $[mm\ day^{-1}]$, and $L$ is the latent heat of evaporation in $[KcalKg^{-1}]$.

The evaporation rate $(E)$ is given by:

$$E = 7.44 \times 10^{-5} u_4(e_s - e_a) \qquad \text{Marciano} - \text{Harbeck formula in metric units} \tag{45}$$

where $u_4$ is the wind speed at $4m$ above water surface $[Kmhr^{-1}]$.

The latent heat of vapourization is given:

$$L = 597.1 - 0.57T_s \tag{46}$$

where $L$ is the latent heat of evaporation in $[KcalKg^{-1}]$

Environmental Laboratory (1986) recommends equation 47 or 48 depending on the value value of $(e_s - e_a)$:

$$\phi_e = -\rho_w L(a + bW)(e_s - e_a) \quad \text{for} \;\; e_s \; \geq e_a \tag{47}$$

where $\phi_e$ is the evaporative losses in $[Kcal\ m^{-2}sec^{-1}]$, $a$ is a empirical coefficient that is associated with the units of wind speed= $0.25 \times 10^{-9}$, and $b$ is a empirical coefficient that is associated with the units of wind speed= $1.0 \times 10^{-9}$.

$$\phi_e = 0 \qquad\qquad \text{for} \;\; e_s \; \leq e_a \tag{48}$$

The vapour pressure of air is given by:

$$e_a = 2.171 \times 10^8 \ exp\left[\frac{-4157}{T_d + 239.09}\right] \tag{49}$$

where $T_d$ is the dew point temperature [deg $C$].

Henderson-Sellers (1986) recommends the following equation (source: Ryan et al., 1974):

$$\phi_e = -\Big[\lambda(T_{wv} - T_{av})^{0.33} + b_* u_2\Big]\Big(e_s - e_a\Big) \tag{50}$$

where $\phi_e$ is the evaporative losses in $[Wm^{-2}]$, $T_v$ is virtual temperature, $\lambda$ is an empirical coefficient $(= 2.7 \times 10^{-2}\ [Wm^{-2}(Nm^{-2})^{-1}K^{-0.33}])$, and $b_*$ is an empirical coefficient

($= 3.2 \times 10^{-2}$ $[Wm^{-2}(Nm^{-2})^{-1}K^{-0.33}]$). The value of $\lambda$ is zero when $T_{wv}$ is less than $T_{av}$.

The virtual temperature is given by

$$T_v = \frac{T}{1 - (0.378\frac{e}{P})} \tag{51}$$

where $P$ the is atmospheric pressure and $e$ is the vapour pressure.

Imberger and Patterson (1981) do not explicitly recommend a specific formula for the evaporative heat losses, but it is implicitly understood that the Environmental Laboratory (1986) recommendation is acceptable.

It is worth noting that Jobsen (1972) stated that errors in the computation of evaporation will be diminished by almost seven times if $3hr$ data are utilized in these computations instead of the customary daily averaged values. Caution is also invoked during spring when the Bowen's ratio is plagued with irregularities as the lake is warming with the subsequent errors in computing the heat content (Bolsenga, 1975). Often in the earlier studies during the 1970's when the evaporation formulae were still being improved, the estimates from the energy budget studies were used to calibrate and verify the empirical constants used in the mass-transfer evaporation formulation technique.

### 2. 3. 5 *Conductive losses ($\phi_c$)*

This is the smallest component of the surface heat fluxes. It is sometimes referred to as the sensible heat flux. The computational formulae for the conductive heat losses incorporate the use of evaporative losses via the Bowen's ratio. The classical Bowen's ratio as defined in 1926 assumed that the bulk transfer coefficient for water vapour and heat are the same. This was lator found to be only under neutrally stable atmospheric conditions. Developments in the 1950's improved the formulation by adding terms that would account for the atmospheric stability conditions.

Livingston and Imboden (1989) recommends the following equation:

$$\phi_c = -C_B f(T_s - T_a) \tag{52}$$

where $C_B$ is the Bowen coefficient (0.65 $[hPaK^{-1}]$ at standard atmospheric pressure), and $f$ is wind function (see equation 43).

Orlob et al. (1983) recommends the following equation:

$$\phi_c = -f(Ri)\ \rho_w\ c_p\ N \frac{P}{P_0}\ u(T_s - T_a) \tag{53}$$

where $c_p$ is the specific heat of air at constant pressure (0.219 $Kcal\ kg^{-1}\ K^{-1}$), $P$ is the atmospheric pressure $[mbar]$, $P_0$ is standard atmospheric pressure at sea-level (1013 $mbar$), $u$ is wind speed in $[Kmhr^{-1}]$, $N$ is the dimensionless evaporation coefficient ($\approx 1.40 \times 10^{-6}$ at $z = 2\,m$ for land/lake data), and $f(Ri)$ is the Richardson number function.

The Richardson number is given by:

$$Ri = \frac{-g(\rho_a - \rho_0)z}{\rho_a u^2} \tag{54}$$

where $z$ is height above water surface, $g$ is acceleration due to gravity, and $\rho_a$ is the density of air at $z = 2m$, and $\rho_0$ is the density of saturated air at $T_s$.

The Richardson number function is given for three conditions of Richardson number as shown below:

$$f(Ri) = (1 - 22Ri)^{0.80} \quad for \;\; 0 \geq Ri \geq -1 \tag{55}$$

$$f(Ri) = (1 + 34Ri)^{-0.80} \quad for \;\; 0 \leq Ri \leq 2 \tag{56}$$

$$f(Ri) = 1 \quad for \;\; Ri = 0 \; (neutral\ case) \tag{57}$$

For the purposes of performing these computations, $\rho_a$ was assumed to be 1.29 $Kg/m^3$ and $\rho_0$ to be 1.30 $Kg/m^3$. This was held constant through-out the computations.

Environmental Laboratory (1986) recommends the following equation:

$$\phi_c = -\rho_w\, L(a + b\, W)\Big(C_B + 1 \times 10^{-3}\, P\Big)\Big(T_s - T_a\Big) \tag{58}$$

where $\phi_c$ is the conductive losses $[Kcal/m^{-2}/sec]$, $P$ is the barometric pressure $[mbar]$, $C_B = 0.61 \deg C^{-1}$ is Bowen's ratio, $W$ is wind speed in $[msec^{-1}]$, and $a\&b$ are empirical coefficients (see equation 47).

Henderson-Sellers (1986) recommends the following equation:

$$\phi_c = -0.61 \times 10^{-3} \phi_e P \frac{\Big(T_s - T_a\Big)}{\Big(e_s - e_a\Big)} \quad \text{for assumed neutral atmospheric stability} \tag{59}$$

where $\phi_c$ is conductive losses $Wm^{-2}$, and $P$ is atmospheric pressure $Nm^{-2}$.

Imberger and Patterson (1981) do not explicitly recommend a specific formula for the conductive heat losses, but it is implicitly understood that the Environmental Laboratory (1986) recommendation is acceptable.

# Chapter 3 Solar Radiation Analysis

## 3.1 Methodology of the Solar Radiation Analysis

The net solar radiation on the water surface ($\phi_{s,actual}$) is the product of the computed solar radiation and the reflection factor, $F_r$. This reflection factor takes into account the albedo effects which is assumed to be a function of Julian day and cloud cover.

The net solar radiation is given by the equation below.

$$\phi_{s,actual} = \phi_s \; F_s \qquad (60)$$

Environmental Laboratory (1982) and Henderson-Sellers (1986) do provide an explicit formulation for $F_s$. Hence it is assumed that the specification in Environmental Laboratory (1982) is satiafactory for the other three methods. The measured solar radiation data is likewise assumed to follow the same specification for $F_s$ as in Environmental Laboratory (1982).

The measured solar radiation values were obtained from unpublished data by the Upstate Freshwater Institute. The measurements were taken at hourly intervals and summed to obtain the averaged daily solar radiation. All except Brock (1981) simulates the hourly radiation values. The hourly values are subsequently added up and divided by 24 hours to obtain the averaged daily solar radiation so that comparisons can be made with the measured values. A FORTRAN program Solar. for (see appendix) performs these computations and comparisons using the r. m. s. statistic. The basic meteorological input data consists of the daily averaged values of atmospheric pressure, wind speed, dry bulb temperature, dew point temperature and cloud cover. These meteorological variables were obtained using the published data from NOAA at Hancock airport.

## 3.2 Results and Discussion

The simulations were optimized for each of the models of solar radiation by varying the specific variables in such a way as to obtain the least root mean square error.

The following optimization procedures were carried out in the computations.

- Krambeck (1982): The thickness of the atmosphere was optimized at $10 Km$.

- Brock (1981): From deJong (1973), two locations namely Blue Hill, Massachusetts and Madison, Wisconsin were selected as substitutes for the climatic constants of Syracuse as no previous study was done for Syracuse. The selection was based on similar latitude and vegetation. It was found that the annual climatic constants of Blue Hill, Massachusetts as shown below produces the least r. m. s. value.

  $$a = 0.22 \ \& \ \ b = 0.50$$

  These values are close to the suggestions by Black, Bonythan, and Prescott (1954), and Prescott (1956).

- Environmental Laboratory (1986): The optimized values of the dust attenuation coefficient, $d$, for each of the simulated years are given in Table 2.

- Environmental Laboratory (1982): The optimized values of the atmospheric turbidity factor, $T$, for each of the simulated years are given in Table 2.

- Henderson-Sellers (1986): As previously mentioned, the atmospheric transmission term of Environmental Laboratory (1986) was equated to the atmospheric transmission term in the Henderson-Sellers (1986) formulation. Hence there was no unique parameter to be independently optimized.

Two separate segments of year 1989 were compared and hence given the connotation $1989-1$ and $1989-2$. Figures $9-16$ illustrates the results of these computations for years 1985, 1986, 1987, 1988, 1889, 1990, and 1991 respectively. The respective r. m. s. values are also shown in Table 2. Based on the r. m. s. values Environmental Laboratory (1982) performs most adequately with a mean r. m. s. value of $33.8 Kcal/m^2/hr$. Figures $17-24$ shows the measured values compared with the best-fit i.e. the Environmental Laboratory (1982) method.

The possible sources of error in the above analysis:

1. The human errors associated with the observation and detection of measurements as well the complicated nature of the maintainance of these equipments. With care this could be minimized and avoided.

2. The practically and conceptually difficult task of measuring the cloud cover. Lund (1968) estimated that amongst all the meteorological variables, cloudcover had the highest negative correlation with the incident solar radiation (-(0.8–0.9)). This was the highest correlation existing between the incident solar radiation and any of the other meteorological parameters. Kasten and Czepak (1979) and Norris (1968) incorporated the type and amount of cloud cover into the solar radiation models.

Table 2: The r. m. s. values for the five solar radiation models

| Yr | A(*) | B(**) | C | D | E | F |
|---|---|---|---|---|---|---|
| 1985 | 31.9(0.3) | 30.8(2.0) | 105.5 | 53.7 | 55.1 | 30.7 |
| 1986 | 41.5(0.2) | 39.1(1.5) | 7 7.4 | 75.9 | 54.7 | 40.1 |
| 1987 | 49.1(0.1) | 33.6(2.0) | 122.5 | 60.0 | 55.4 | 33.6 |
| 1988 | 35.6(0.4) | 34.5(2.5) | 101.8 | 50.3 | 70.6 | 34.8 |
| 1989-1 | 35.1(0.4) | 36.5(2.5) | 92.0 | 47.4 | 72.9 | 38.9 |
| 1989-2 | 20.1(0.7) | 17.2(1.5) | 170.5 | 49.8 | 31.4 | 19.9 |
| 1990 | 38.6(0.4) | 37.7(2.5) | 116.7 | 51.0 | 68.1 | 38.7 |
| 1991 | 40.6(0.3) | 40.7(2.5) | 67.2 | 71.8 | 71.6 | 40.4 |
| Average r.m.s. | 36.6 | 33.8 | 106.7 | 57.5 | 59.9 | 34.6 |

*Values of d
**values of T

A: Environmental Laboratory (1986)
B: Environmental Laboratory (1982)
C: Krambeck (1982)
D: Brock (1981)
E: Henderson Sellers (1986)
F: Environmental Laboratory (1982) (for T=2.1)

**NOTE:** all values in $Kcal/m^2/hr$

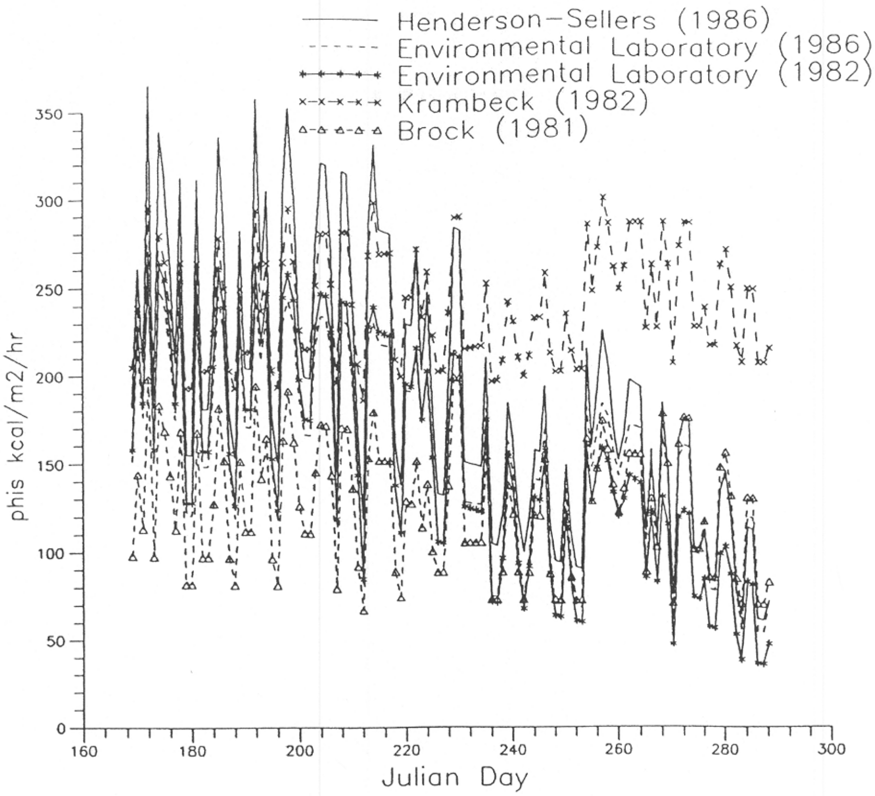


Figure 9: Solar radiation computations from the five models for 1985

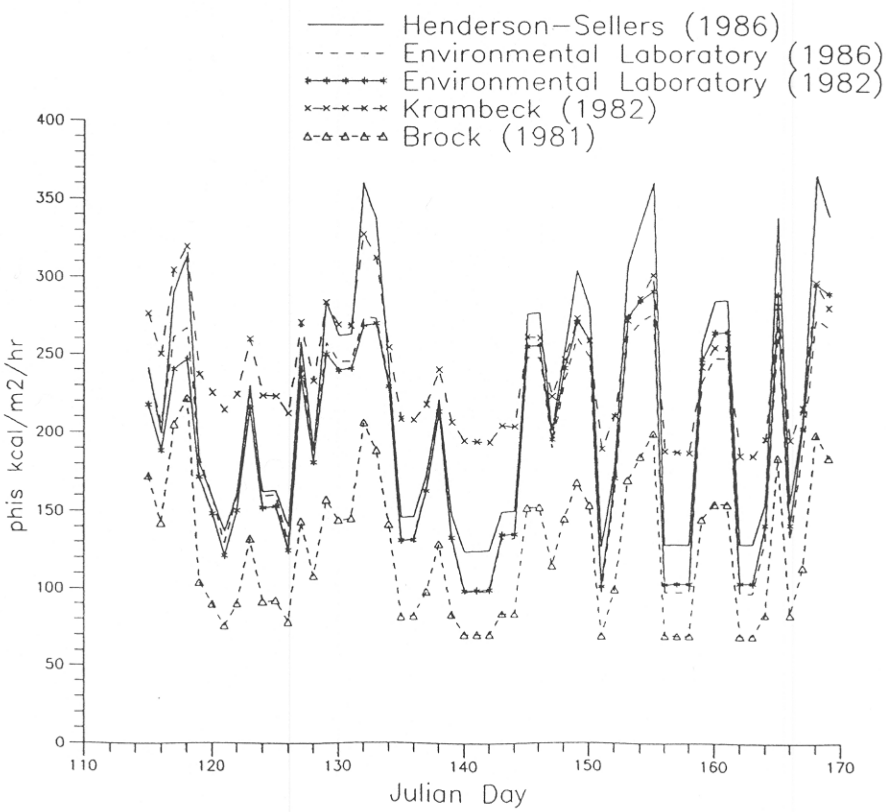


Figure 10: Solar radiation computations from the five models for 1986

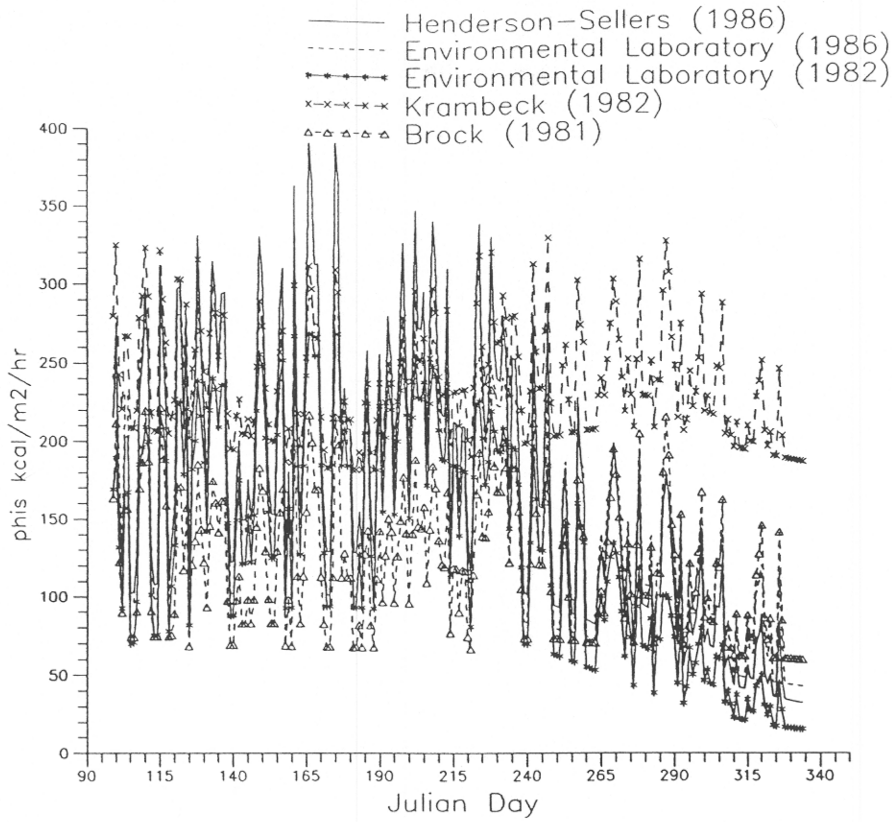


Figure 11: Solar radiation computations from the five models for 1987

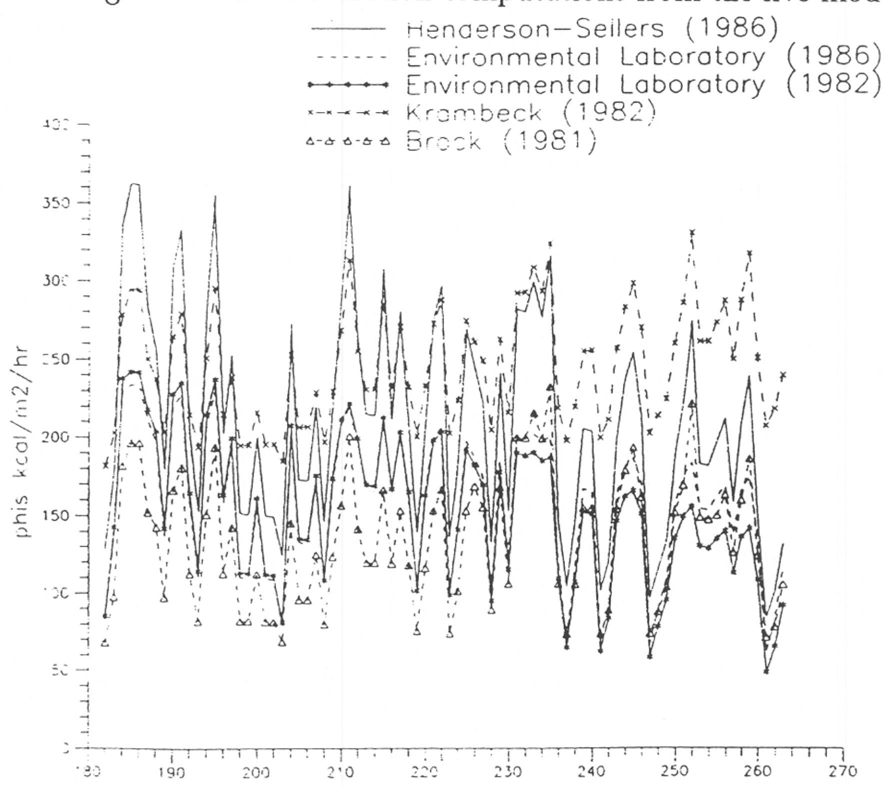


Figure 12: Solar radiation computations from the five models for 1988

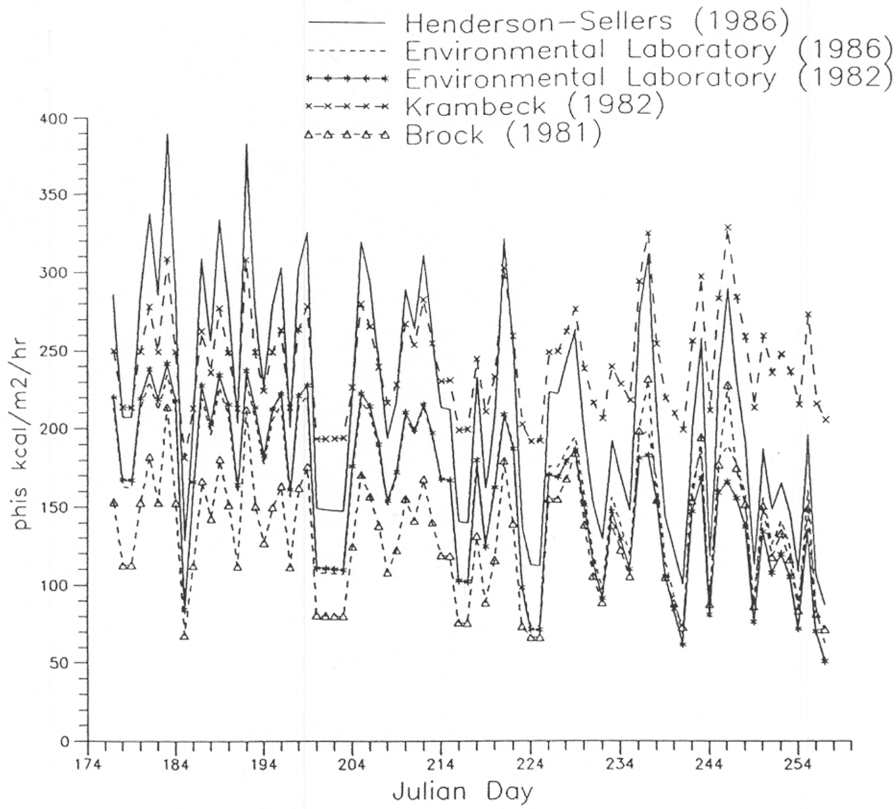


Figure 13: Solar radiation computations from the five models for 1989-1

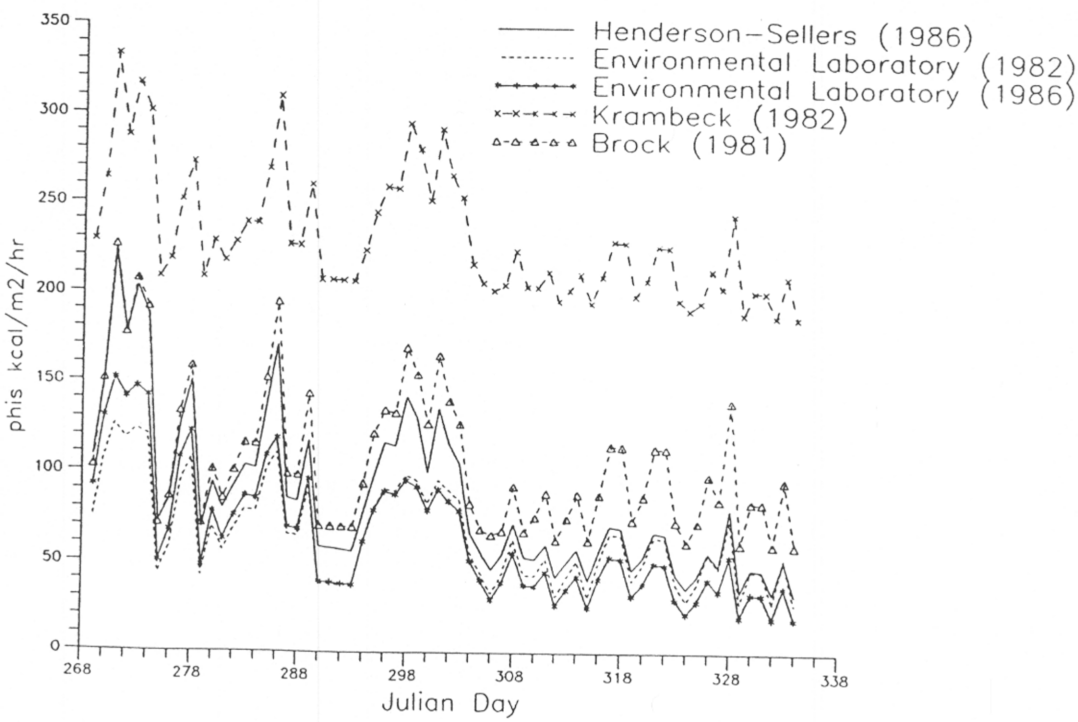


Figure 14: Solar radiation computations from the five models for 1989-2

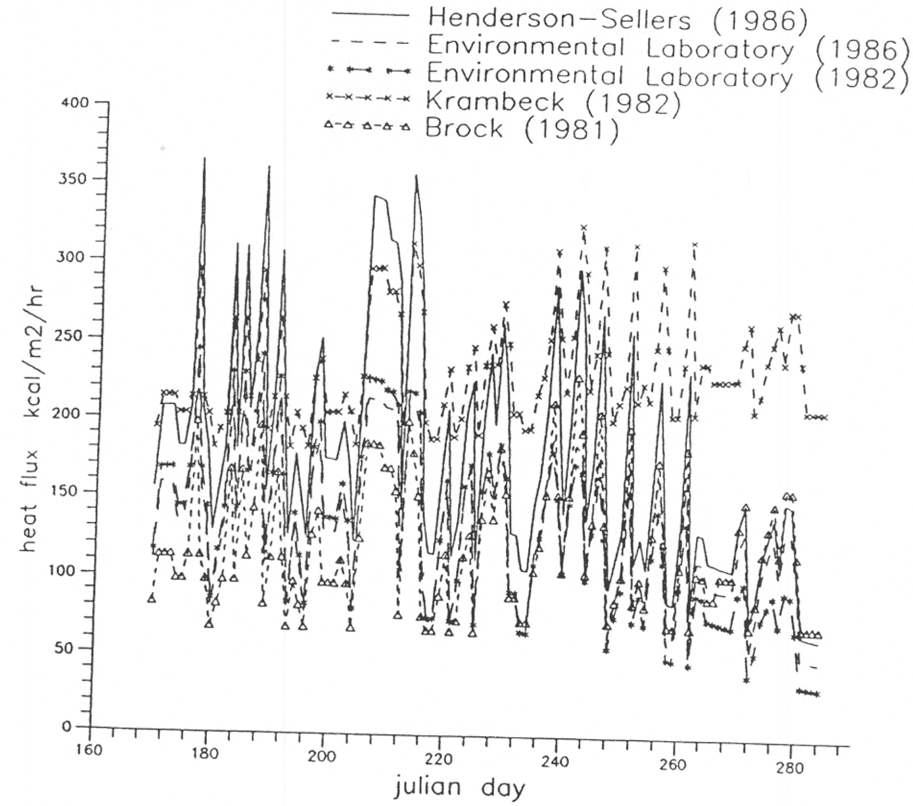


Figure 15: Solar radiation computations from the five models for 1990

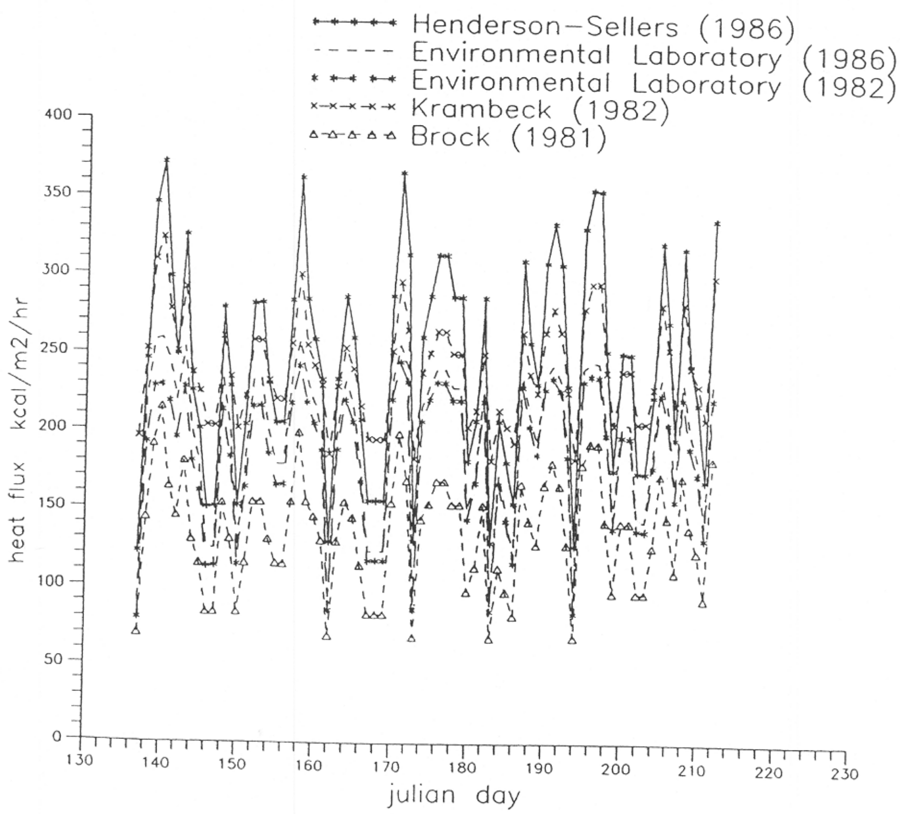


Figure 16: Solar radiation computations from the five models for 1991

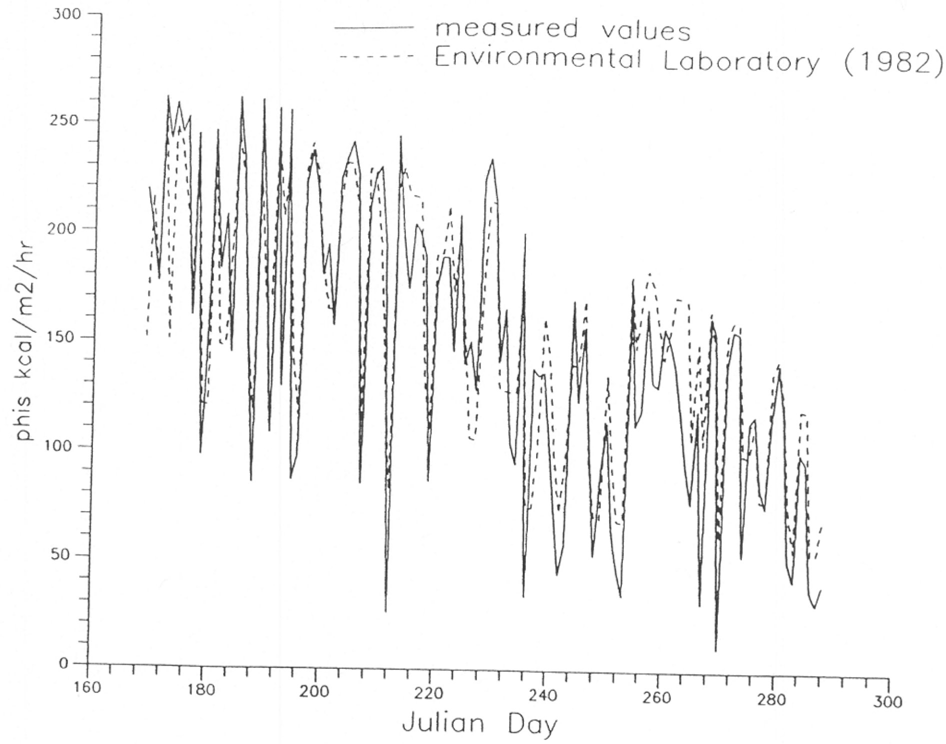


Figure 17: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1985

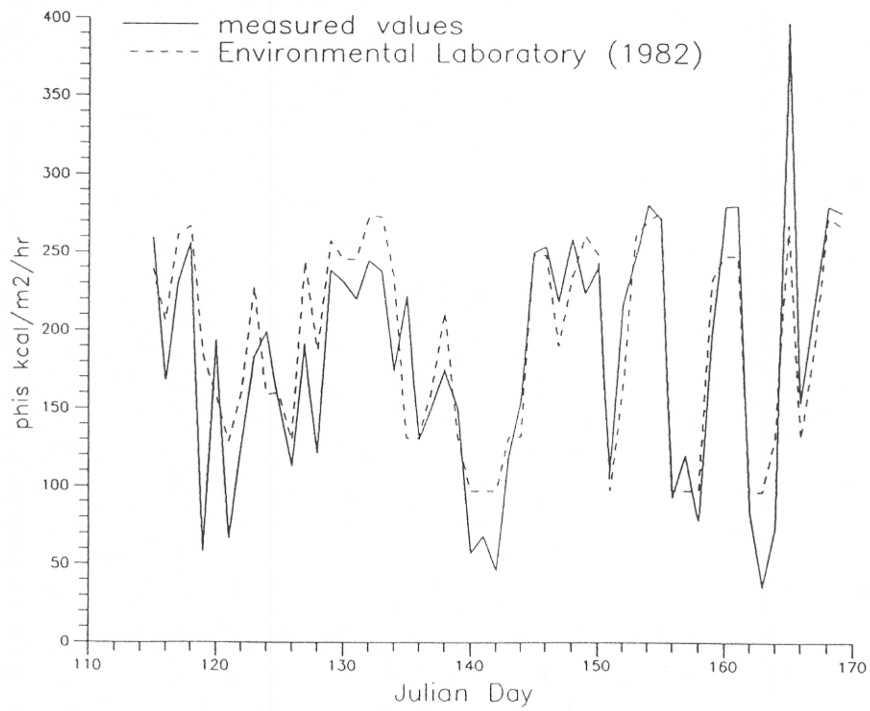


Figure 18: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1986

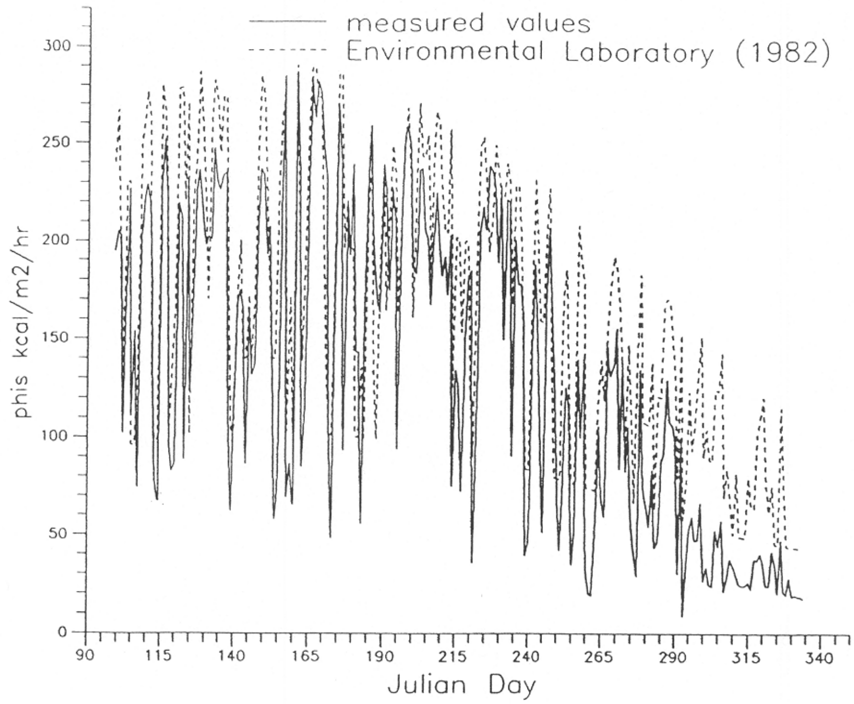


Figure 19: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1987

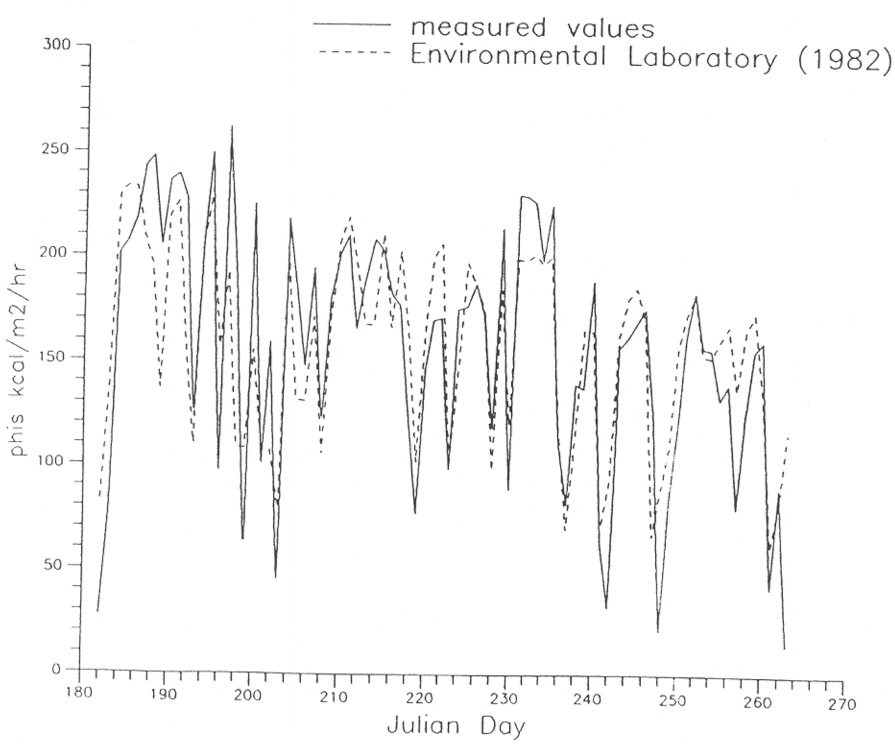


Figure 20: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1988

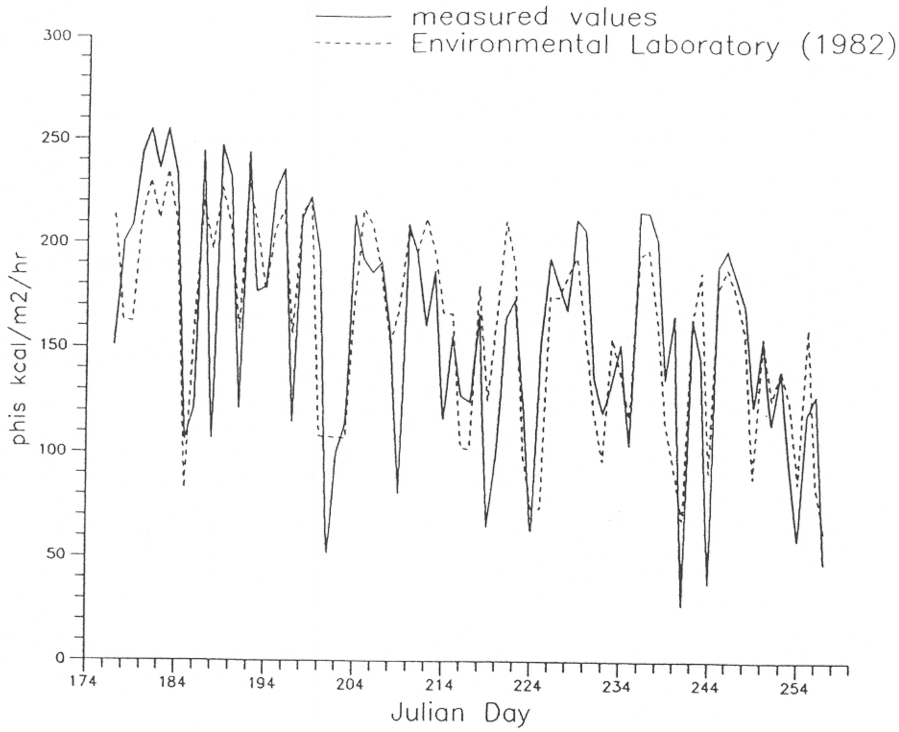


Figure 21: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1989-1

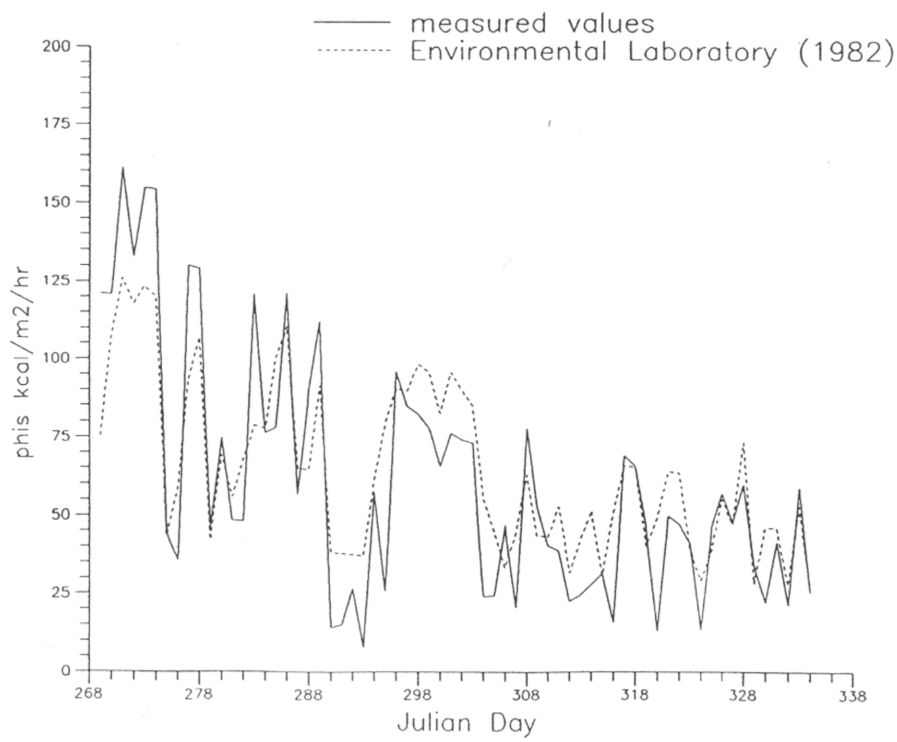


Figure 22: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1989-2

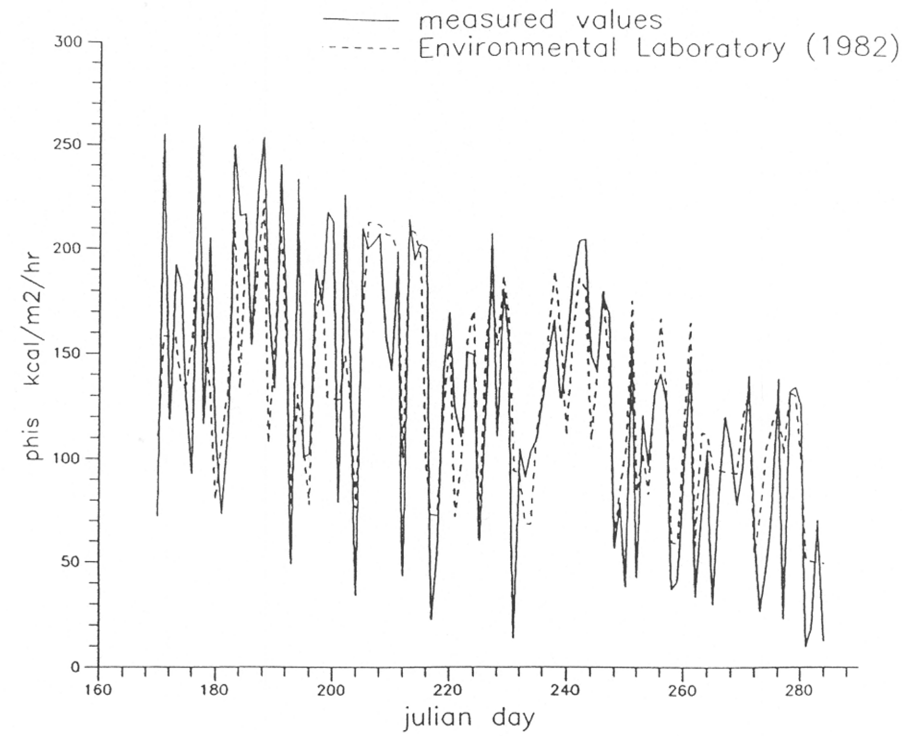


Figure 23: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1990

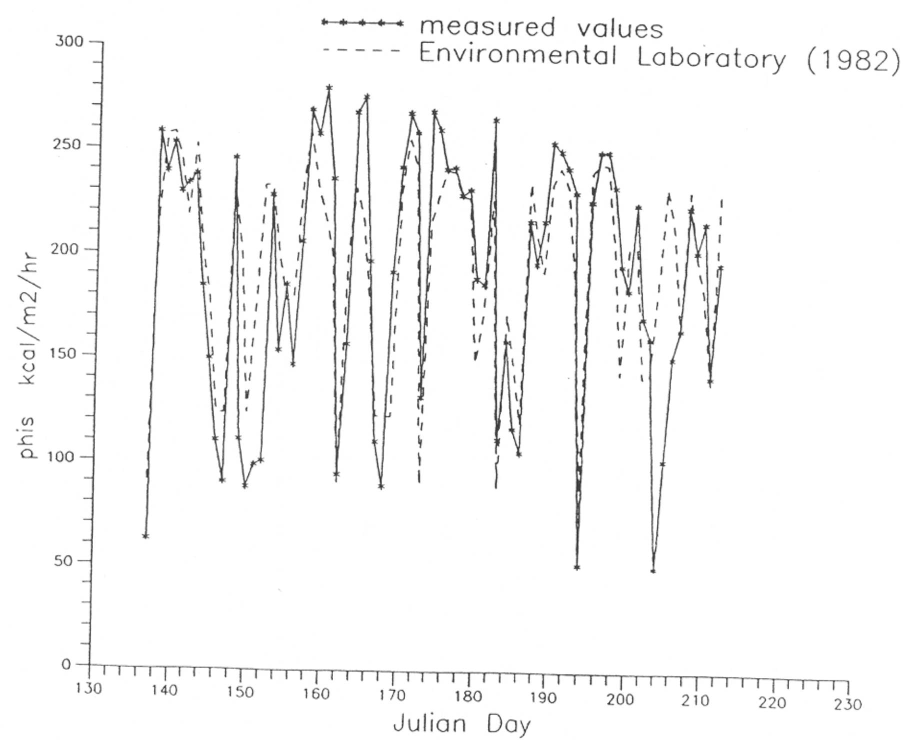


Figure 24: Solar radiation values using Environmental Laboratory (1982) and the measured values for 1991

In Stigter (1982), the uncertainties associated with average point cloudiness are attributed to the variations in the cloud type and amounts during the day, and the nunber of cloud observations made.

3. In Bolsenga (1975), Richards and Lowen (1965) are quoted as having as having developed a water to land cloud ratio with values ranging from 1.37 (October) to 0.89 (April). This was attributed to the fact that more cumulus clouds were developed over land than lakes in late spring, summer, and early fall, while during late fall and winter, more cumulus and stratocumulus clouds developed over the lake than over the land (Bolsenga, 1975).

These limitations must be borne in mind before giving a qualitative interpretation to the results of these models.

## 3.4 Recommendations on the Solar Radiation Models

The following observations and conclusions can be deduced from the results presented in this chapter.

- Both Environmental Laboratory models perform adequately and the numerical difference in the respective r. m. s. values cannot be said to be significant. But based on this study for Onondaga Lake, Environmental Laboratory (1982) method is recommended.

- Both Brock (1981) and Henderson-Sellers (1986) come up second with almost no significant numerical difference between the two. They have nearly twice the r. m. s. values as the first-comers. But based on the simplicity of formulation and meage need for meteorological input variables, Brock (1981) performed surprisingly well and hence would be recommended as a viable alternative for preliminiary studies or in the absence of weather data.

- Krambeck (1982) has almost twice the mean r. m. s. values as the second-comers or four times the first-comers. But the formulation is unique and simple, and comparatively is the most deterministic of all the models. It was recently developed and therfore refinements and adaptability would naturally be expected.

- For prognostic modeling purposes, Environmental Laboratory (1982) is recommended with a suggested mean value of turbidity factor of 2.1 for Onondaga Lake. This was the average of all the optimized turbidity factors shown in Table 2. A sample run was created using this value for all the years of study and the r. m. s. values are shown in Table 2. The results are convincing and the numerical difference between the individually optimized Environmental Laboratory (1982) and at $T = 2.1$ is insignificant.

# Chapter 4 Net Surface Heat Flux Analysis

## 4.1 Methodology Utilized in the Surface Heat Flux Analysis

### 4.1.1 *Computation of measured net surface heat flux* ($\phi_{N,meas.}$)

The net surface heat flux is the time derivative of the heat storage within the lake. Any changes in the heat content has a direct reflection on the prevailing surface heat fluxes. The total heat content of the lake is given by

$$\textsf{total heat in lake} = \rho c T_{ave} V \tag{61}$$

where $V$ is the total volume of the lake and $T_{ave}$ is the averaged lake temperature.

Using the principles of heat conservation, the rate of change of heat content of a lake can be expressed as:

$$\frac{d}{dt}(\rho c T_{ave} V) = \rho c Q_{IN} T_{IN} - \rho c Q_{OUT} T_{OUT} - \phi_N A_{SURF} \tag{62}$$

where $Q_{IN}$ and $Q_{OUT}$ are the inflow and outflow discharges, $T_{IN}$ and $T_{OUT}$ are the inflow and outflow temperatures, and $A_{SURF}$ is the surface area of the lake.

The following assumptions are taken into account in the formulation of $\phi_{N,meas.}$:

- The ground-water flow into the lake is considered negligible.
- Flow of lake water into the ground-water system is also considered nil.
- The lake's sediment-water boundry is considered insulated .
- The precipitation/evaporation is assumed not to contribute to the change in lake volume.

- The temperatures of the inflows and outflows are considered equal.

The net surface heat flux ($\phi_N$) given by equation 62 can be simplified using the assumptions given above to result in the equation given below.

$$\phi_{N,meas.} = \rho \; c \frac{V}{A_{SURF.}} \frac{\Delta T_{ave}}{\Delta_{time}} \tag{63}$$

$\Delta_{time}$ is the sampling interval between profile measurements (days), and $\Delta T_{ave}$ is the difference in the averaged lake temperature on consecutive days of profile measurements.

The temperature profile measurements for Onondaga Lake were obtained from unpublished data from the Upstate Freshwater Institute. The measurements were taken roughly about every week during the latter stages of the morning hours. Each profile measurement were taken at regular intervals of 1 m till the maximun depth of 19 m was reached.

The procedure for the computation of $\phi_{N,meas.}$ consists of two main steps. They are:

1. On days of temperature profile measurement, the averaged lake temperature is computed using the lake morphometric equations as given below.

   The surface area (A) of the lake at any elevation is given by

$$A = a_1 \left[ (Z_s)^{a_2} \right] \tag{64}$$

   where $A$ is area at any elevation $Z_s$ (maximum $Z_s$ for Onondaga lake is $19m$), $a_1$ is a constant ($1.408 \times 10^6$), and $a_2$ is a constant (0.701).

The volume (V) enclosed by the elevation is given by

$$V = \left(\frac{a_1}{a_2 + 1}\right)\left[Z_s^{a_2+1}\right] \tag{65}$$

where $V$ is volume enclosed by the elevation $Z_s$.

2. The net surface heat flux $\phi_{N,meas.}$ between consecutive days of temperature profile measurements is computed using equation 63.

A FORTRAN program Thesis1.for (see Appendix) was written to perform these computations.

4.1.2 *Formulation of the computed net surface heat flux* ($\phi_{N,comp.}$)

$$\phi_{N,comp.} = \phi_a + \phi_s - \phi_e - \phi_c - \phi_b \tag{66}$$

where

$\phi_{N,meas.}$ =computed net surface heat flux

$\phi_a$ =atmospheric heat flux

$\phi_s$ =incident solar radiation flux after reflection

$\phi_e$ =evaporative heat flux

$\phi_c$ =conductive heat flux

$\phi_b$ =back radiation

Equation 66 described above is utilized in evaluating the $\phi_{N,comp.}$ by each of the five recommendations for net surface heat fluxes.

A FORTRAN program 22thesis.for (see appendix) performs these iterations in addition to the evaluation of the r. m. s. statistic as described below:

$$r.m.s. = \sqrt{\frac{\left(\sum_{i=1}^{n-1}\left(\phi_{N,meas.} - \phi_{N,comp.}\right)^2\right)}{n-1}} \tag{67}$$

where

$\phi_{N,meas.}$ =measured net surface heat flux

$\phi_{N,comp.}$ =computed net surface heat flux

$n$ =number of temperature profile measurements during the year.

The basic meteorological inputs to the 22thesis.for program mentioned above are the wind speed, atmospheric pressure, dry bulb temperature, dew point temperature, and cloud cover. It is worth noting that the use of land station data for temperature, wind speed, and vapour pressure as opposed to the lake data, in the computation of $\phi_e$, $\phi_c$ and $\phi_a$ can contribute to errors (Bolsenga, 1975).

## 4.3 Results and Remarks

Figures 25-31 illustrates the computed net surface heat flux as evaluated by the five recommendations for years 1985, 1986, 1987, 1988, 1989, 1990 and 1991 respectively. The computed r. m. s. values for the respective five recommendations are given in Table 3. Figures 32-38 provide a close-up view between the measured net surface heat flux and the best-fit of the computed net surface heat flux. Based on the r. m. s. statistic, Livingston and Imboden (1989) performed the best with an average r. m. s. value of 82.3 $Kcal/m^2/hr$.

Table 3: The r. m. s. values for the five net heat flux models

| Yr | A | B | C | D | E |
|---|---|---|---|---|---|
| 1985 | 73.8 | 134.3 | 70.2 | 64.8 | 101.4 |
| 1986 | 47.4 | 94.5 | 69.9 | 51.1 | 80.8 |
| 1987 | 112.3 | 120.8 | 126.9 | 114.9 | 118.5 |
| 1988 | 78.9 | 68.8 | 118.2 | 74.9 | 65.6 |
| 1990 | 98.3 | 107.3 | 106 | 102.5 | 106.5 |
| 1991 | 96.5 | 101.9 | 110.7 | 96.6 | 103.8 |
| Average r.m.s. | 83.4 | 101.1 | 100.8 | 82.3 | 92.1 |

A: Environmental Laboratory (1986)
B: Orlob et.al. (1983)
C: Imberger and Patterson (1981)
D: Livingston and Imboden (1989)
E: Henderson Sellers (1986)

**NOTE:** all values in $Kcal/m^2/hr$

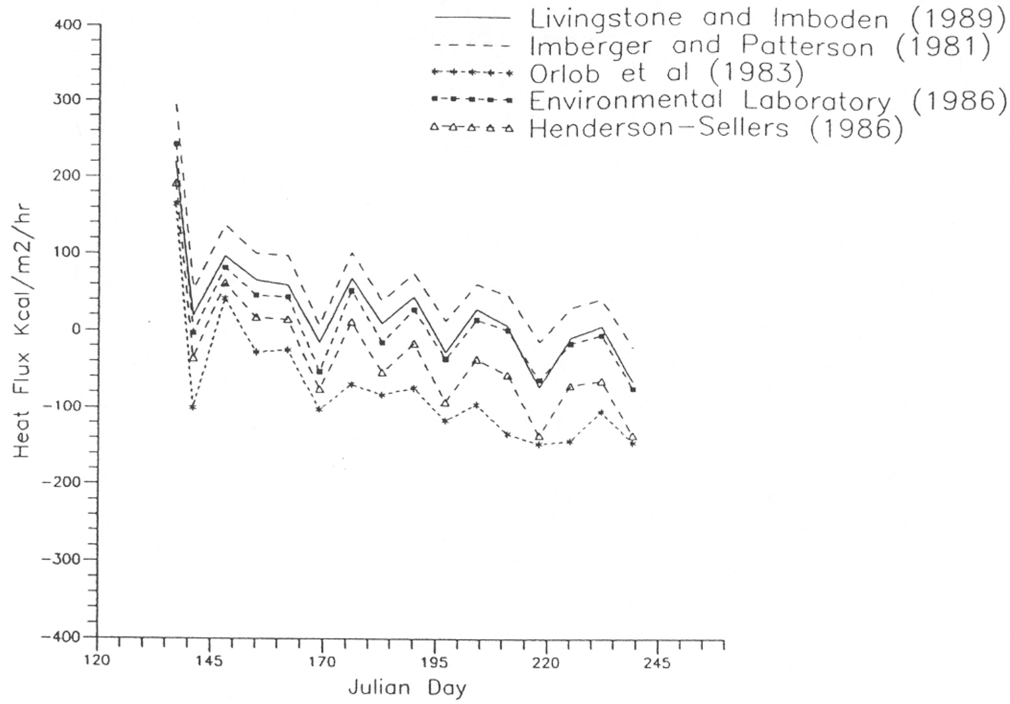


Figure 25: The net surface heat flux as computed by the five models for 1985

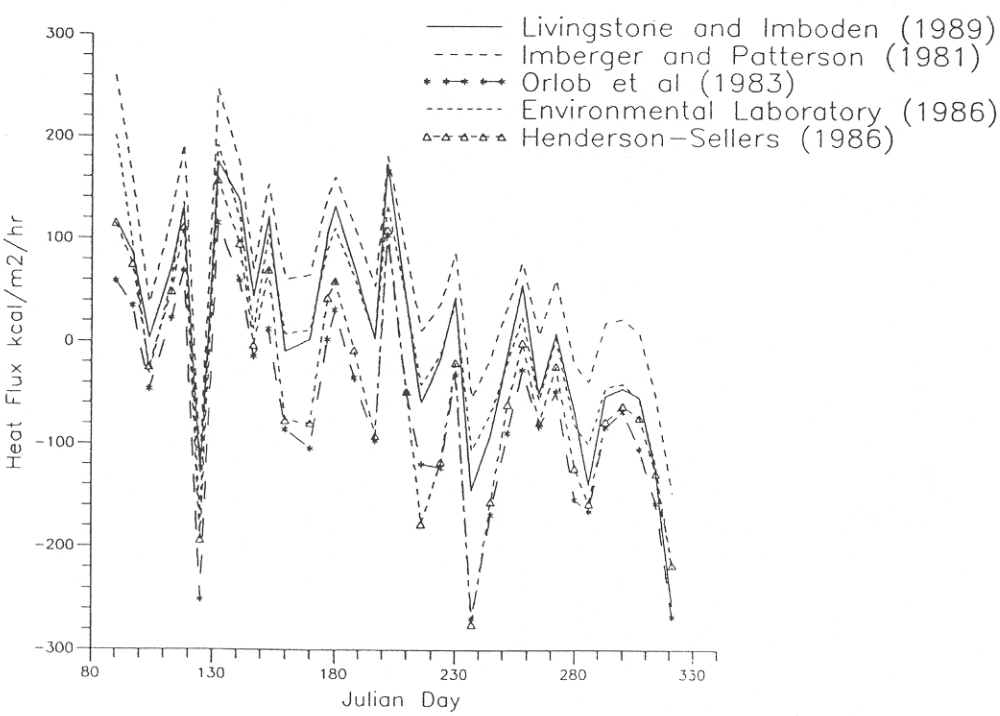


Figure 26: The net surface heat flux as computed by the five models for 1986

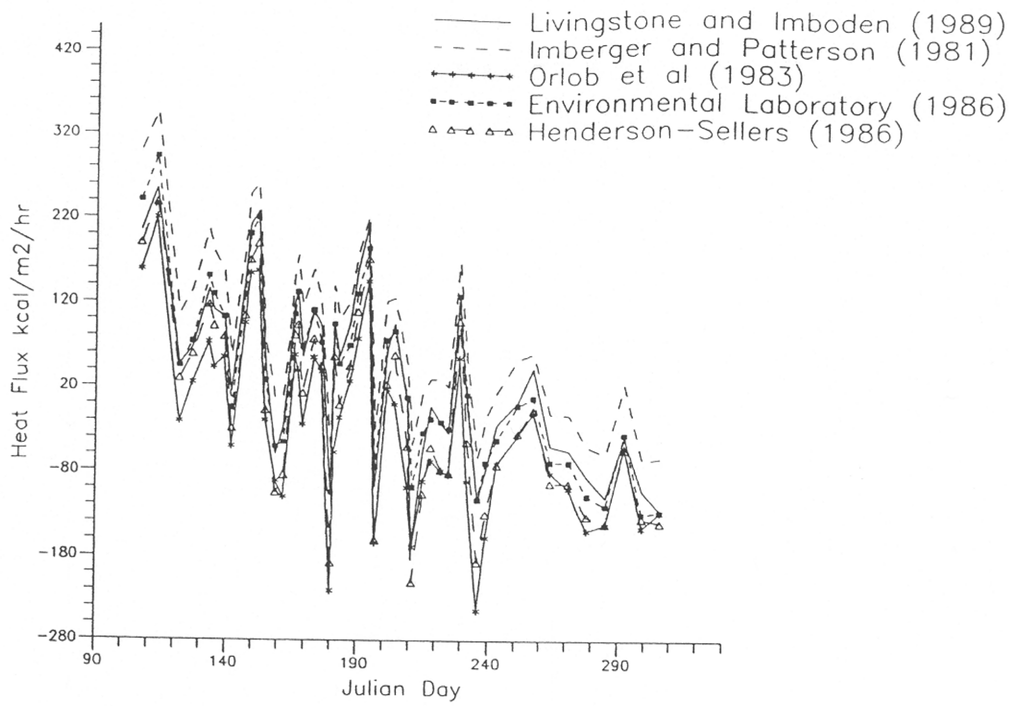


Figure 27: The net surface heat flux as computed by the five models for 1987

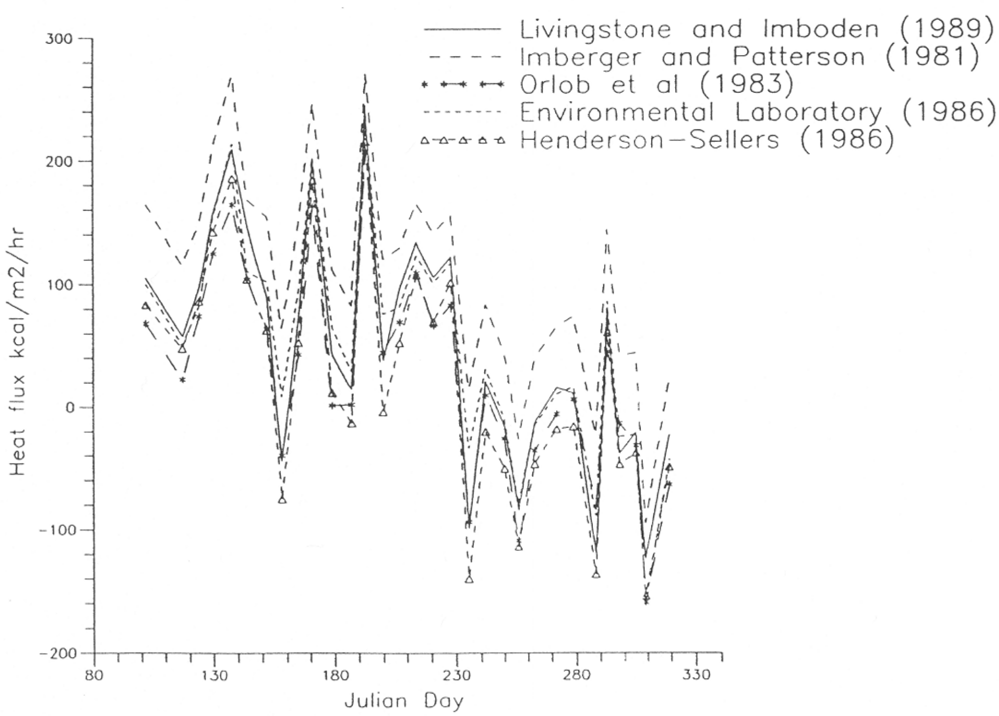


Figure 28: The net surface heat flux as computed by the five models for 1988

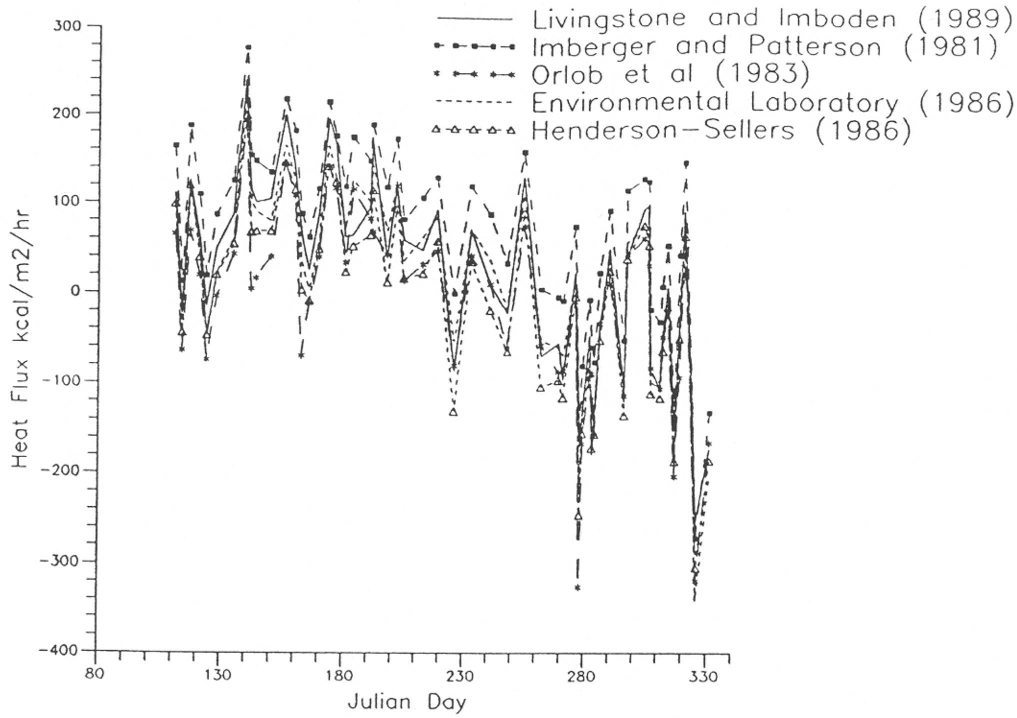


Figure 29: The net surface heat flux as computed by the five models for 1989

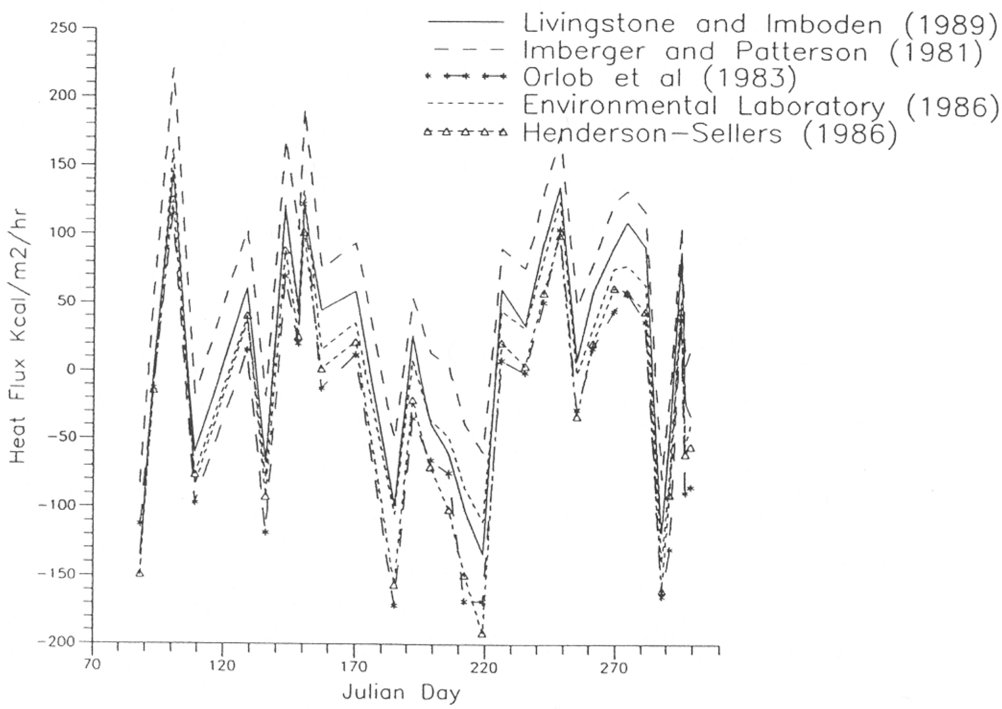


Figure 30: The net surface heat flux as computed by the five models for 1990

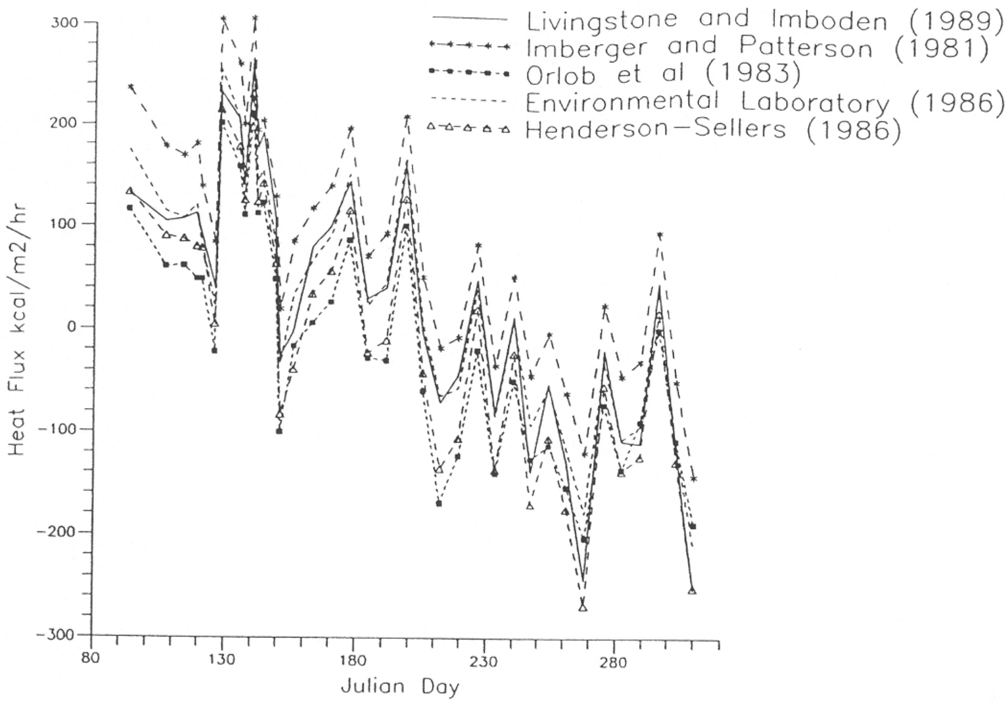


Figure 31: The net surface heat flux as computed by the five models for 1991

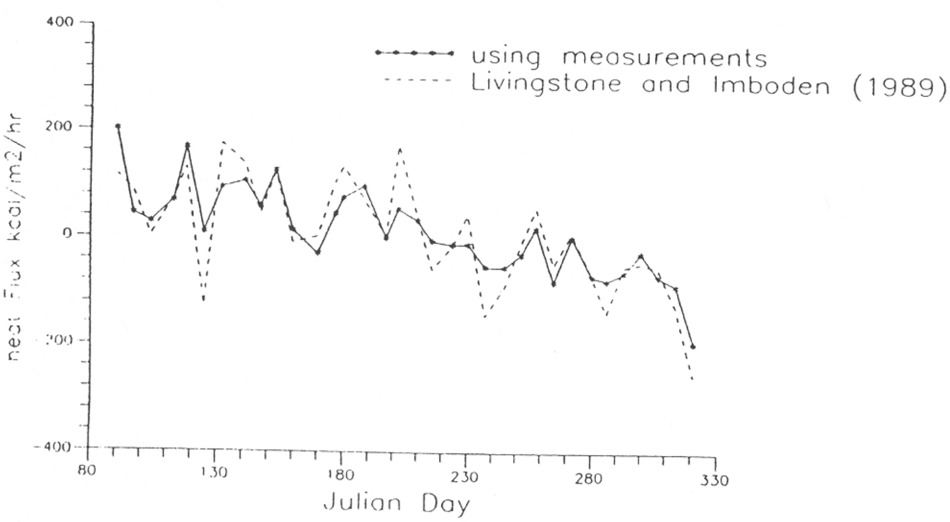


Figure 32: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1985

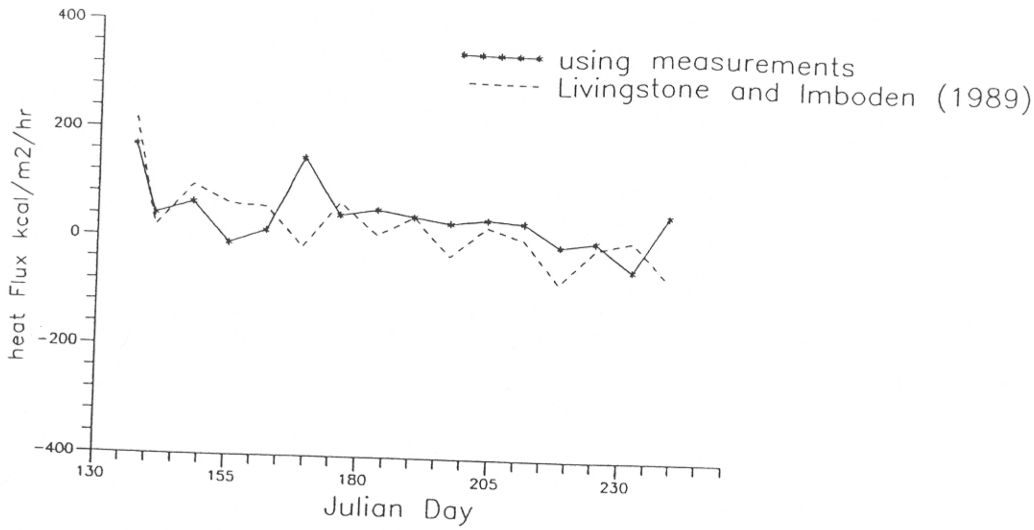


Figure 33: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1986

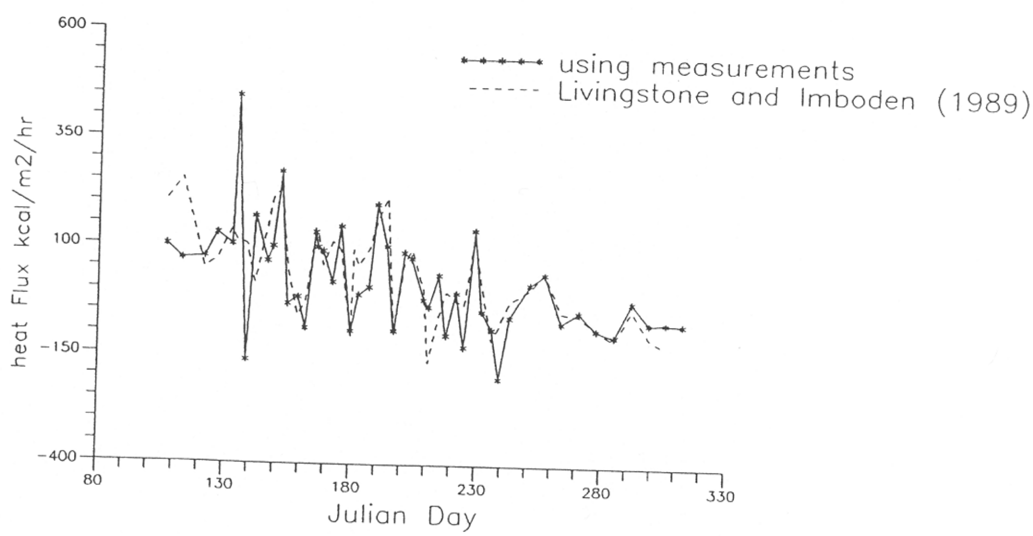


Figure 34: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1987

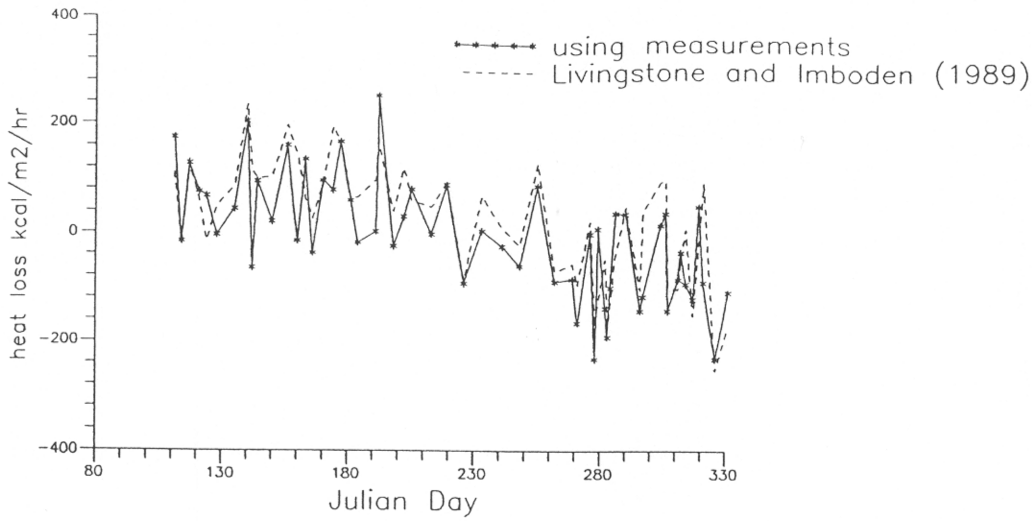


Figure 35: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1988

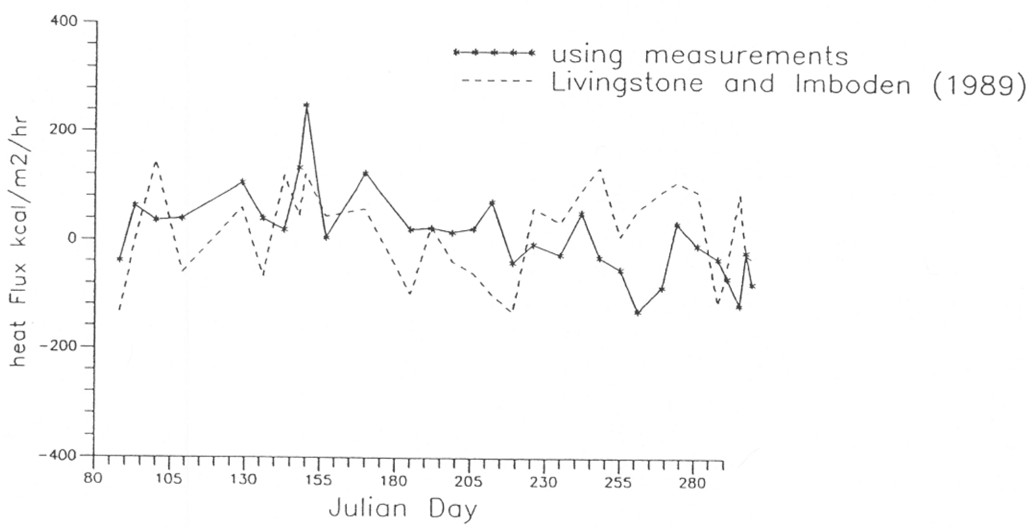


Figure 36: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1989

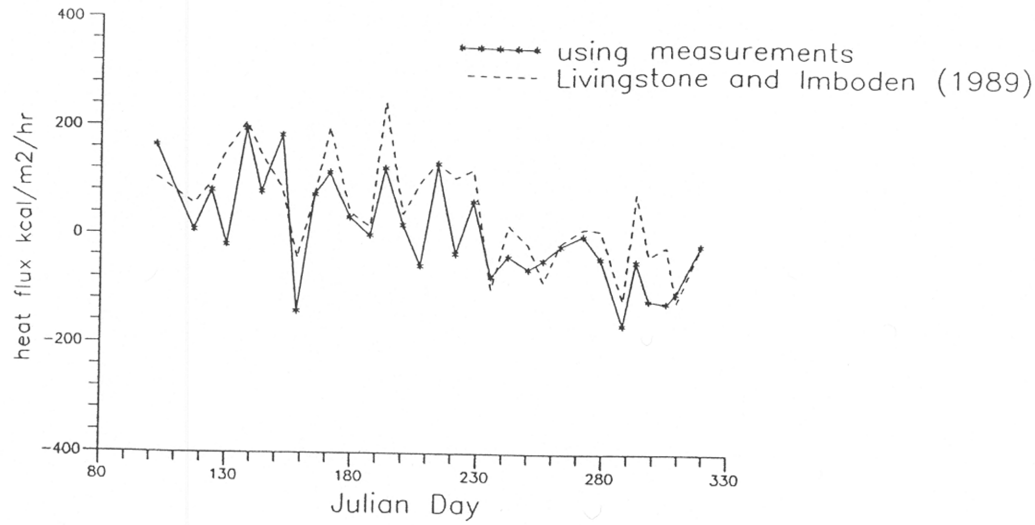


Figure 37: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1990

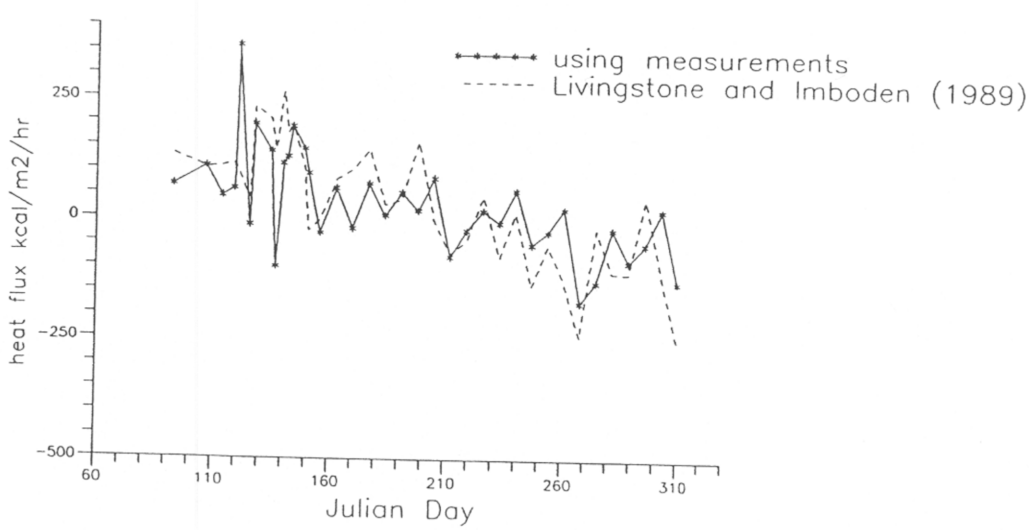


Figure 38: The net surface heat flux as computed by Livingstone and Imboden (1989) and the measured values for 1991

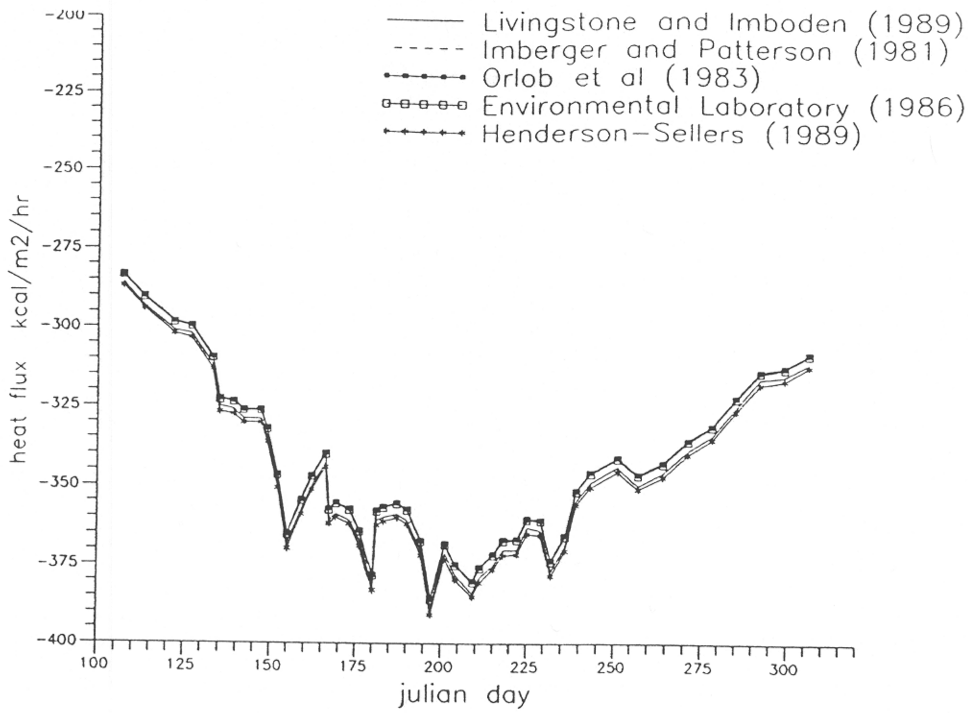


Figure 39: The computed back radiation values using the five recommendations for 1987

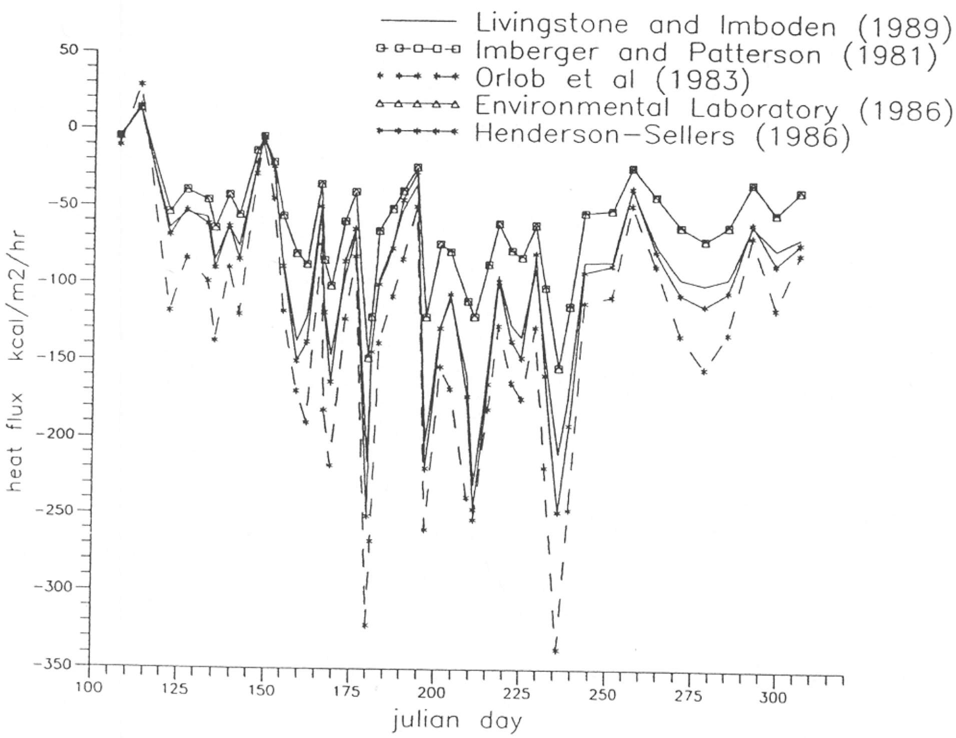


Figure 40: The computed evaporative heat fluxes using the five recommendations for 1987

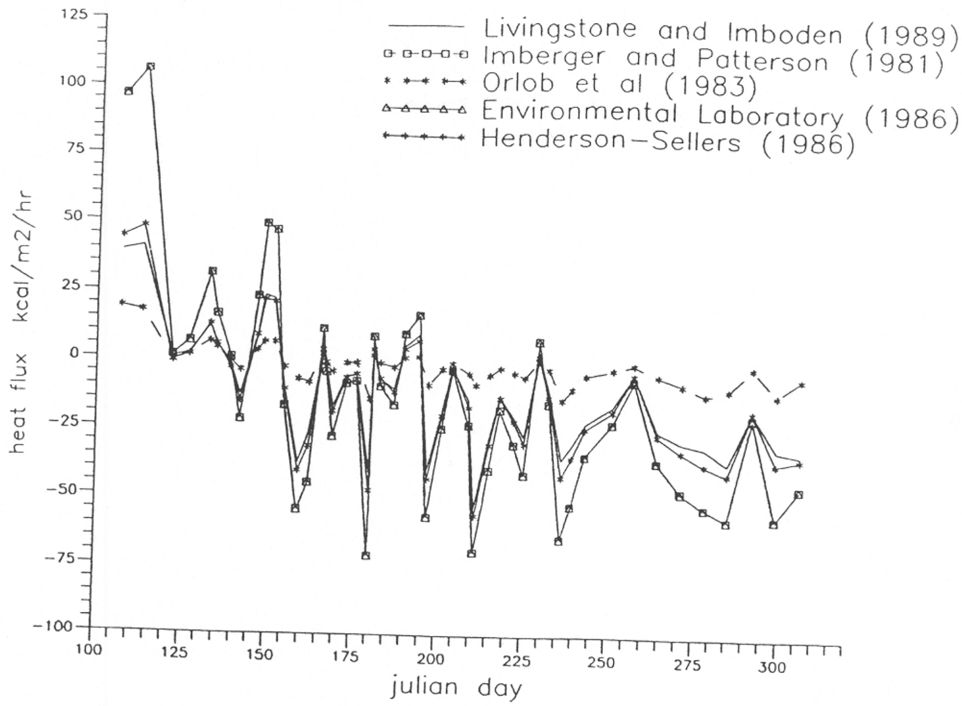


Figure 41: The computed conductive heat fluxes using the five recommendations for 1987

Based on the results presented above, the following deductions can be made:

Livingston and Imboden (1989) is the best-fit of all the recommendations. But the difference between it and Environmental Laboratory (1986) can be said to be insignificant. Therefore this study would seem to validate the use of Environmental Laboratory (1986) in CE-QUAL-R1 (1986).

Henderson-Sellers (1986) came second with an average r. m. s. value 11% higher than the best-fits described above.

Orlob et al. (1983) and Imberger and Patterson (1981) came last with an average r. m. s. value 20% higher than the best-fits. But the difference between them can be considered insignificant.

Figure 39 illustrates the computed back radiation for year 1987 by the five respective recommendations. As earlier mentioned, there is a universal agreement as to the

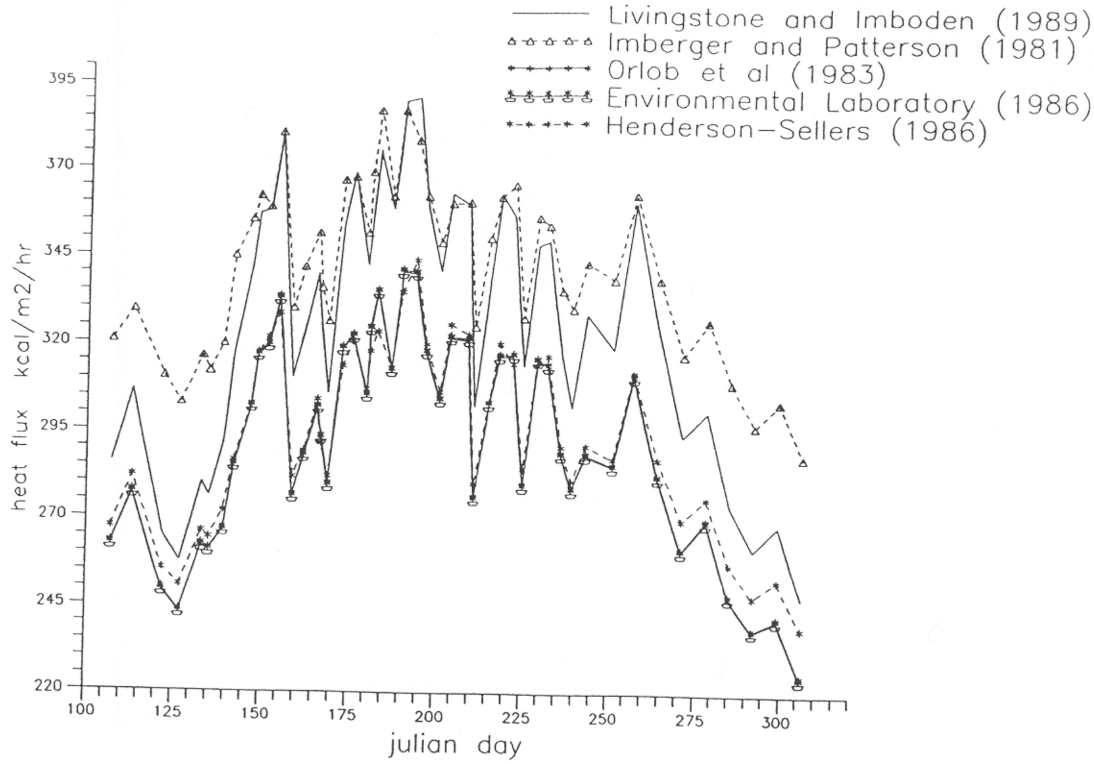


Figure 42: The computed atmospheric radiation fluxes using the five recommendations for 1987

nature of its formulation.

Figure 40 illustrates the computed evaporative heat fluxes for year 1987 by the five respective recommendations. Environmental Laboratory (1986) and Imberger and Patterson (1981) are the same due to the earlier mentioned assumption. Orlob et al. (1983) has the greatest magnitude with the Henderson-Sellers (1986) closely following second. Livingston and Imboden (1989) hugs closely to the latter with only a small difference. Environmental Laboratory (1986) and Imberger and Patterson (1981) have the least magnitude of all and closely follows the Livingstone and Imboden (1989) formulation. The differences in these recommendations can be attributed to the site-specific nature of these formulations with respect to the meteorological variables at Syracuse.

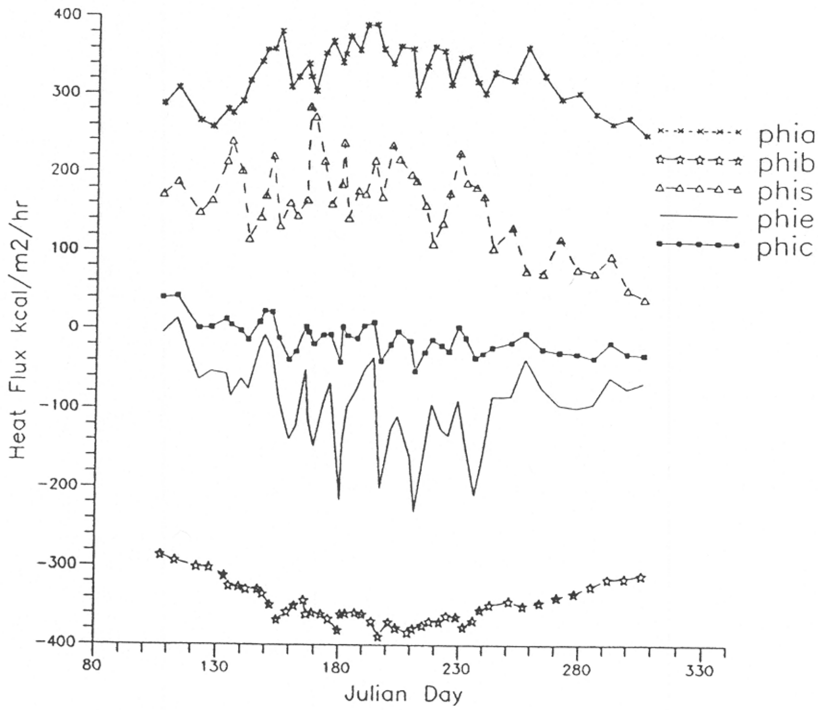


Figure 43: The net surface heat flux and its components using Livingstone and Imboden (1989) for Onondaga Lake in 1987

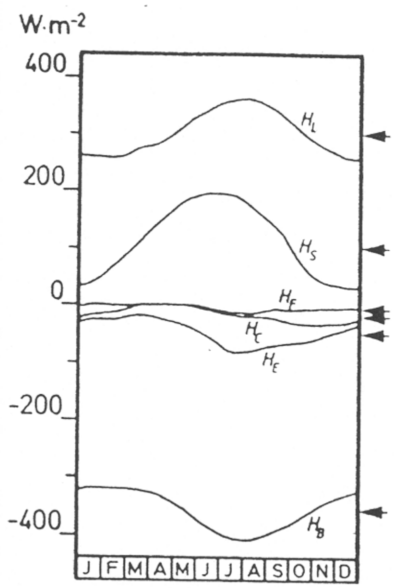


Figure 44: The net surface heat flux and its components for Lake Aegeri as in Livingstone and Imboden (1989)

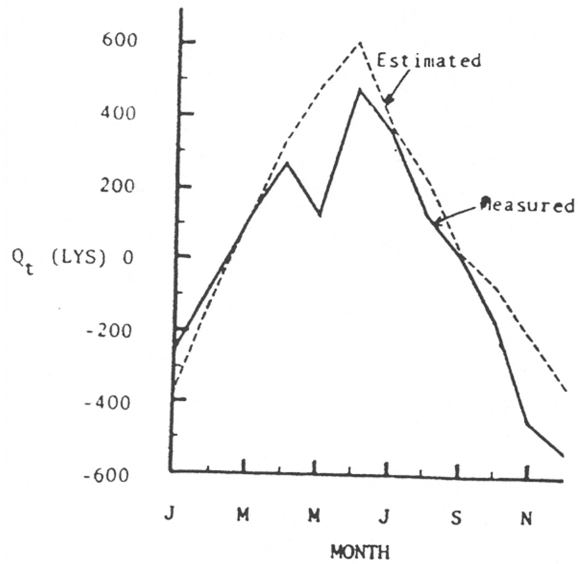


Figure 45: The net surface heat flux as measured from bathythermograph data and as computed from energy budget methods for Lake Huron. Source: Bolsenga, 1975

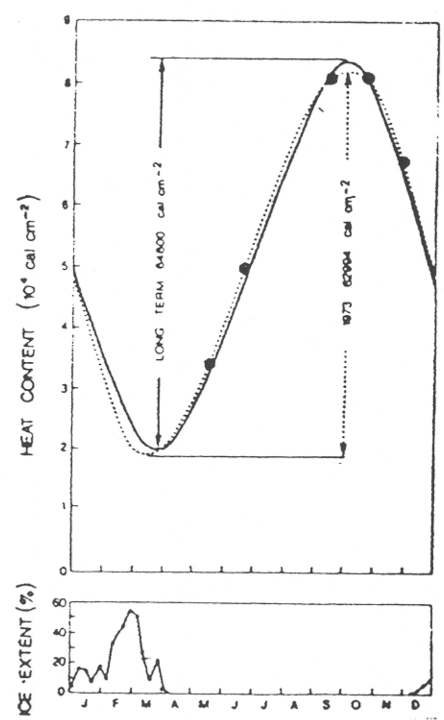


Figure 46: The long-term change in heat content for Lake Superior and the actual change for 1973 Source: Schertzer, 1978

Figure 41 illustrates the computed conductive heat fluxes for year 1987 by the five respective recommendations. Environmental Laboratory (1986) and Imberger and Patterson (1981) are the same due to the previously mentioned assumption. Environmental Laboratory (1986) and Imberger and Patterson (1981) shows the greatest magnitude with the Henderson-Sellers (1986) closely following second. Livingston and Imboden (1989) hugs closely to the latter with only a small difference. Orlob et al. (1983) shows more or less a uniformly small magnitude.

Figure 42 illustrates the computed atmospheric radiation for year 1987 by the five respective recommendations. Environmental Laboratory (1986) and Orlob et al. (1983) apparently have the same formulation. Imberger and Patterson (1981) shows the greatest magnitude with Imboden and Livingston (1989) closely following second. Henderson-Sellers (1986) closely follows next with Orlob et al. (1983) hugging it closely.

Figure 43 illustrates the different components of the net surface heat flux during the year 1987 using the best-fit i. e. Livingston and Imboden (1989) for Onondaga Lake. The longwave radiations which includes the back radiation and atmospheric radiation are the biggest components of the surface heat flux and seem to cancel each other to a large degree with no apparent phase lag between them. This is similar to the trend shown in Figure 44 for Lake Aegeri, Switzerland.

Figure 45 for Lake Huron in Bolsenga (1975) is presented for comparison with Figures 33-39. The measured values in May are reportedly flawed due to logistical difficulties. It utilized monthly averaged values of the meteorological variables unlike this study.

Figure 46 for Lake Superior in Schertzer (1978) also is presented for comparison. The figure shows the change in the heat content of Lake Superior as opposed to the surface heat flux this study dealt with. It utilizes averaged monthly weather values and it shows a sinusoidal variation in the heat content.

# Chapter 5 CONCLUSIONS AND RECOMMENDATIONS

Water quality modeling is essentially an art of duplicating the behaviour of environmental systems by interfacing the relevant and often, interdependent physical, biological and chemical processes using theoretical, experimental and empirical observations as guiding tools. They are based on the principles of conservation of mass and heat and the advection-diffusion concepts to define the hydrodynamic conditions. Most models use the one-dimensional vertically averaged, layered model structure to simulate the system behaviour. However this basis should be based on the model objectives, prevailing processes and mechanisms in the lake, and the lake morphometry. In order to simulate the episodal events, the temporal and spatial scales used in the simulation must be smaller than the respective scales associated with the episodal event (Ford and Thornton, 1979). Most environmental data are spatially and temporally dependent. It could be suggested that the use of 3-hr intervals, should be sufficient enough even for the most detailed work. The surface heat fluxes which this study dealt with in detail is an essential component to the accurate understanding and therefore the predictive ability of most water quality models. The results of this study can also be used for the design of solar ponds, aquaculture ponds and in the climatological study of global warming.

In conclusion, Environmental Laboratory (1982) model for solar radiation is suggested as a recommendation for future studies of Onondaga Lake. Livingstone and Imboden (1989) is suggested as the best recommendation for the rest of the surface flux components.

The following recommendations are made for any future study of Onondaga lake:

1. The annual heat budget study should be made by extending the present study to the months when the lake is covered with ice.
2. A proper study into the best possible combination for the prediction of the atmospheric, evaporative, conductive and back-radiation formulae using indirect evidences and empirical evidences and the results of this study.

```
c       thesis1.for
c       This program looks at the real heat balance of Onondaga Lake
c       as evaluated from the temperature profile measurements at
.       specific intervals of time during the course ofthe year.
        e -the evaporation/precipitation term is ignored in the
c       current anlysis.
c       Also ignore inflow/outflow terms for the mean time
c       The average lake temperature is averaged as on hand-out
c       Input units used are as follows:
c       Temp[deg.Celcius],Time[days],Area[m2],Volume[m3],density[Kg/m3],
c       & spe.heat.capacity(c) - [Kcal/Kg/deg Celius]


        DIMENSION AvT(200),DTEM(200),DELT(200),A(50),V(50),actvol(50)
       <,phiN(100),T(60,80),depth(60,80),Julian(200),N(80),
       <sumTaV(80),S(60,80),dep(60,80)
        real K,J

        open (unit=6,file='thesi91_out.dat')
        open (unit=10,file='91.dat',status='old')

        NN=39
        delz=0.5
        c=1
        rho=1000
        a1=1.408E6
        a2=0.701
        Zs=19.5



c       calculates A and V as a function of depth
        DO 300 J=1,NN
        IF(J.EQ.2)go to 18
        IF(J.EQ.1)go to 17
        IF(J.GT.2)go to 21

17      A(J)=a1*((Zs)**a2)
        V(J)=(((a1)/(a2+1))*(Zs**(a2+1)))
        go to 300


18      Zs=Zs-0.25

        A(J)=a1*((Zs)**a2)
        V(J)=(((a1)/(a2+1))*(Zs**(a2+1)))

        go to 300

21      Zs=Zs-delz

        A(J)=a1*((Zs)**a2)
        V(J)=(((a1)/(a2+1))*(Zs**(a2+1)))


  00    continue

c       calculates the actual volume of each segment after initialising
c       the volume enclosed by the lake bottom as zero
        DO 333 J=1,NN
```

```
      V(40)=0.00
      actvol(J)=V(J)-V(J+1)
 333  continue

      DO 400 K=1,38




32    read(10,*)Julian(K),N(K)


      DO 900 J=1,N(K)
48    read(10,*)Depth(J,K),T(J,K)
      dep(J,K)=19.5-depth(J,K)
900   continue
400   continue
      totV=(((a1)/(a2+1))*(19.5**(a2+1)))
      AS=a1*((19.5)**a2)


      DO 600 K=1,38
      sumTaV(K)=0.0

c     On certain Julian daysof certain calender years,the measurements
c     were carried up to one
c     or two layers less than the complete.But on the average,the
c     bottom and the one or two layers above it will almost the same
      average temperature-this is an assumption onlyused in the coding.

69    DO 700 J=1,N(K)


62    sumTaV(K)=sumTaV(K)+T(J,K)*actvol(J)

700   continue
      go to 01

01    AvT(K)=sumTaV(K)/totV

600   continue

      AS=a1*((19.5)**a2)

      DO 222 K=2,38

      DTEM(K)=AvT(K)-AvT(K-1)

      DELT(K)=(Julian(K)-Julian(K-1))*24

      phiN(K)=((rho*c*TotV)/(DELT(K)*AS))*DTEM(K)

      WRITE(6,98)Julian(k),PHIN(K),AvT(K),DTEM(K),DELT(K)
98    format(5(E12.5,3x))
  2   continue
      stop
      end
```

```
c     solar.for
c     Solar radiation is calculated using the set of equations as provided in
c     CE-QUAL-R1(new), CE-QUAL-R1(old), Krambeck (1982), Brock et al. (1981)
      and Henderson-Sellers (1986),
c     and compared with the measured values to draw up a statistical
c     conclusion.In addition, it calculates the r.m.s value for
c     CE-QUAL-R1 (old) at T=2.19.
      DIMENSION cloud(600),dbt(600),dpt(600),apress(600),ws(600)
     <,phis(600),delta(600),alpha(600),phiscalcu(600),Aphiscalcu
     <(600),junk0(500),junk(500),Aphiscalc(600)
     <,Fdpt(600),phiscalc(600),Rphis(600),albedo(600)
     <,RR(600),Y(600),del(600),gamma(600),ENE(600)
     <,Dum(600),AENE(600),nice(600)
     <,aene1(600)
     <,aene2(600),ene2(600)
     <,phisprog(600),Aphisprog(600)

      real omega,oam,phis,ALT,d,Q0,z,zz,zzz,zzzz,delta,tL,
     <fc,R,sum,alpha,A,B,fr,a1,a11,fs,phiscalcu,Rphis
     <,Aphiscalcu,phi,Fdpt,dpt,phiscalc,Aphiscalc,PPP
     <,K,L,nice
     <,NN,W1
     <,kkk
      integer tsu,tss



      open (6,file='sol91_out.dat')
      open (11,file='w91_24hr.dat')
      open (12,file='inc91.dat')



      phi=((2*3.142)*42.9)/(360)
      ALT=700
      d=0.2
      Q0=1188.0
      T=2.0


      R0=1.495E8
      ee=0.017
      YA=3.1677
      SC=1353
      latti=0.7485
      del0=0.4093
      aa=0.125122
      Re=6.37E3
      thick=10
      delthi=0.5
      nos=thick/delthi
      L=0.0

      aaaa=0.22
      bbbb=0.50


      S=1365
```

```
      kkk=0.33

 2    DO 93 I=86,334

       read(11,33)cloud(I),dbt(I)
     <,dpt(I),apress(I),ws(I)
33     format(18x,f6.2,x,f7.1,x,f7.1,x,f8.1,x,f6.1)
       Fdpt(I)=(dpt(I)*(1.8))+32

93    continue

12    DO 193 I=137,212

      read(12,201)junk1,junk2,phis(I),junk3,junk4,junk5,junk6
     <,junk7,junk8,junk9,junk10
201   format(A9,I4,x,I4,x,I4,x,I4,x,I4,x,I4,x,I4,x,I4,x
     <,I4,x,I4)

193   continue
      S5SD=0.0
      S4SD=0.0
      S3SD=0.0
      S2SD=0.0
      S1SD=0.0
      SSD=0.0
      DO 91 I=137,212

       z=((2*3.142)/(365))*(172-I)
       delta(I)=0.4092*(COS(z))
       Aalbe=0.0
       tL=1.31/15
       zz=((-SIN(phi)*SIN(delta(I)))/(COS(phi)*COS(delta(I))))
       tss=(12/3.14)*ACOS(zz)+tL+12
       tsu=-tss+2*tL+24
       fc=1-0.65*(cloud(I)**2)
       zzzz=(((2*3.14159)/(365))*(186-I))
       R=1+0.17*COS(zzzz)
       diff=tss-tsu
       sum=0
       sum1=0.0
       sum2=0.0
       sum3=0.0
       sum4=0.0

       Y(I)=((I-1)*0.9890109)*0.01745
       w3=sin(del0)*cos(Y(I)-YA)
```

```
      del(I)=ASIN(w3)
      www=Y(I)-YA
      RR(I)=((R0*(1-(ee*ee)))/(1-(ee*cos(www))))

      coef=1-cloud(I)
      NN=(360*I)/(365)*0.01745
      R1=((1)/((1+0.033*cos(NN))**0.5))
      W1=((tss-tsu)/2)*15*0.01745


      CCC=(3.14*(I-81))/(183)
      a0=0.02+0.01*(0.5-cloud(I))*(1-sin(CCC))





c          write(6,2289)I,ccc,a0
c2289      format(3x,3(E12.5,3x))


       DO 92 J=tsu,tss


          omega=(3.14/12)*(J-tL-12)

          x=((SIN(phi)*SIN(delta(I)))+(COS(phi)*COS(delta(I))
     <    *COS(omega)))
          IF(x.le.0.01)x=0.01
          alpha(J)=ASIN(x)

          IF (cloud(I).lt.0.05)then
55        A=1.18
66        B=-0.77

          ELSEIF(cloud(I).lt.0.5.and.cloud(I).ge.0.05)then

56        A=2.20
67        B=-0.97

          ELSEIF(cloud(I).lt.0.95.and.cloud(I).ge.0.5)then
57        A=0.95
77        B=-0.75
          ELSE
58        A=0.35
87        B=-0.45
          ENDIF


88        xx=(57.3*x)
          xx=ABS(xx)

          albedo(J)=A*((xx)**(B))
          IF(albedo(J).ge.1.00)albedo(J)=1.0


          fr=1-albedo(J)
```

```
        zzz=180*alpha(J)/3.14159
        zzz=ABS(zzz)

        oam=(((2.71828)**(-(ALT/2532)))/(SIN(alpha(J))+(0.15*((zzz)
   <    **(-1.253)
   <    ))))




        a1=2.71828**(-(0.465+0.0408*(0.00614*((2.71)**(0.0489
   <    *Fdpt(I)))))
   <    *((0.129+(0.171*((2.718)**(-0.88*oam))))*oam))

        a11=2.71828**(-(0.465+0.0408*(0.00614*((2.71)**(0.0489
   <    *Fdpt(I))))
   <    )*((0.179+(0.421*((2.718)**(-0.721*oam))))*oam))



        PPP=alog10(x)




        f2s=2.71**(-((2.19*(0.128-0.054*PPP))/(x)))

        f1s=2.71**(-((T*(0.128-0.054*PPP))/(x)))
        fs=((a11+0.5*(1-a1-d))/(1-0.5*albedo(J)*(1-a1+d)))
        WWWWW=(15*J)*0.01745

        w33=sin(latti)*sin(w3)-cos(latti)*cos(w3)*cos(WWWWW)
        IF (w33.le.0)w33=0.01
        gamma(J)=asin(w33)


          L=0.0

          DO 4444 K=1,nos
          delD=delthi
          Dum(K)=((exp(-aa*thick)*(Re+thick))/((((Re**2)*(w33**2))
   <             +(2*Re*thick)+(thick**2))**0.5))

          nice(K)=aa*Dum(K)*delD
          L=L+nice(K)


4444      continue


        AAS=((a0)/(a0+sin(alpha(J))))
        phisd=S*sin(alpha(J))*(a11**oam)*0.86004
        phiss=0.38*(S*0.86004-phisd)*sin(alpha(J))
        ene2(J)=(phiss+phisd)*(1-AAS)*(1-(1-kkk)*cloud(I))
```

```
      ENE(J)=fr*SC*((R0/RR(I))**2)*w33*exp(-cloud(I)*L)*0.86004
      phiscalcu(J)=(((fr*fc*fs)*(Q0*(alpha(J))))/(R**2))
      phiscalc(J)=(((fr*fc*f1s)*(Q0*SIN(alpha(J)))))
      phisprog(J)=(((fr*fc*f2s)*(Q0*SIN(alpha(J)))))

      Aalbe=Aalbe+albedo(J)
      sum=sum+phiscalcu(J)
      sum1=sum1+phiscalc(J)
      sum2=sum2+ENE(J)
      sum3=sum3+ene2(J)
      sum4=sum4+phisprog(J)
c     write(6,2289)J,w33,gamma(J),L,ENE(J)
c2289 format(3x,5(E12.5,3x))

92    continue


      AAalbe=Aalbe/diff
      fr1=1-AAalbe

      eneinfi1=(24/3.1415)*(SC/((R1)**2))*((W1*sin(latti)*sin(del(I)))
     < +(sin(W1)*cos(latti)*cos(del(I))))*0.035835

      aene1(I)=fr1*eneinfi1*(aaaa+coef*bbbb)


      Rphis(I)=phis(I)*fr1
      Aphiscalc(I)=sum1/24
      Aphiscalcu(I)=sum/24
      Aphisprog(I)=sum/24

      AENE(I)=sum2/24
      aene2(I)=sum3/24

      DDIFF1=ABS(Rphis(I)-Aphiscalc(I))
      DDIFF2=ABS(Rphis(I)-AENE(I))
      DDIFF=ABS(Rphis(I)-Aphiscalcu(I))
      DDIFF3=ABS(Rphis(I)-aene1(I))
      DDIFF4=ABS(Rphis(I)-aene2(I))
      DDIFF5=ABS(Rphis(I)-Aphisprog(I))

      SD=DDIFF**2
      SD1=DDIFF1**2
      S1SD=S1SD+SD1
      SSD=SSD+SD
      SD2=DDIFF2**2
      S2SD=S2SD+SD2
      SD3=DDIFF3**2
      S3SD=S3SD+SD3
      SD4=DDIFF4**2
      S4SD=S4SD+SD4
```

```
        SD5=DDIFF5**2
        S5SD=S5SD+SD5


        write(6,2289)I,Rphis(I)
     < ,Aphiscalcu(I),SSD,
     < Aphiscalc(I)
     < ,S1SD,AENE(I),S2SD,aene1(I),S3SD,aene2(I),S4SD,Aphisprog(I),S5SD
2289    format(3x,14(E12.5,3x))
91    continue
      stop

      end
```

```
c       22thesis.for
c       This program evaluates the heat loss from the Lake surface
c       based on the recommendations of the four shortlisted
c       journals/documents used in the Onondaga Lake Study.
c       This program precedes the earlier program which evaluated
c       the heat loss from the Lake surface based completely on
c       real measurements.
c       It is with the purpose of making a comparative study that this
c       program was made.
c       The surface layer temp. is assumed to follow the best fit obtained
c       from historical data.


        DIMENSION Ts(60,450),cloud(450),dbt(450),dpt(450)
       <,Iphia(450),phis(450),apress(450),ws(450),TsKEL(60,450),
       <Iphib(450),Iphic(450),Iphie(450),EL(450),f(450)
       <,ea(450),es(450),deldayc(450),Iphin(450),daycc(450)
       <,dayc(450),N(450),IIphia(450),Ophia(450),Qphia(450)
       <,IIphib(450),Ophib(450),Qphib(450),dbtKEL(450)
       <,IIphie(450),Ophie(450),Qphie(450),fRi(450),Ri(455)
       <,IIphic(450),Ophic(450),Qphic(450),Oes(450),Oea(450)
       <,IIphin(450),Ophin(450),Qphin(450),OE(450),Iws(450)
       <,Rphin(450),ISSD(450),IISSD(450),QSSD(450),OSSD(450)
       <,Sphib(450),Sphic(450),Sphia(450),Sphie(450),Sphin(450),SSSD(450)



        real Isigma,IIsigma,Osigma,Iws,
       <Iphie,Iphic,IIphic,IIphia,IIphib,Qphie
       <,IIphie,Iphin,IIphin,fRi,Qphic
       <,ISD,IISD,OSD,QSD,Ri,denair,denairTs,z
       <,ISSD,IISSD,OSSD,QSSD,Rphin,dpt
       <,Sphib,Sphic,Sphia,Sphie,Sphin,SSSD,CR,CD,aa1,Sesw


        open (10,file='w87_24hr.dat')
        open (11,file='onont87.dat')
        open (6,file='221the87s_out.dat')
        open (7,file='222the87s_out.dat')
        open (8,file='223the87s_out.dat')
        open (9,file='224the87s_out.dat')
        open (12,file='thesis1_out.dat')
        open (13,file='sol87_out.dat')
        open (14,file='225the87s_out.dat')


        Cp=0.219
        CB=0.61
        On=1.54E-6
        P0=1013
        Qsigma=4.83E-8
        IIsigma=2.0411E-7
        Osigma=2.041E-4
        Isigma=56.7E-9
        Cat=0.938E-5
        a=0.25E-9
        b=1E-9
        start1=99
        z=5
        denair=1.29
```

```
      denairTs=1.30


      Ssigma=5.6697E-8
      Alambda=2.7E-2
      bstar=3.2E-2



1     DO 90 I=99,334
      read(10,30)cloud(I),dbt(I)
     <,dpt(I),apress(I),ws(I)
     <,phis(I)
30    format(19x,f5.2,x,f7.1,x,f7.1,x,f8.1,x,f6.1
     <,x,f8.1)


c      apress(I)=(1013/29.93)*apress(I)
90    continue



101   DO 91 I=99,334
      read(13,*)er1,phis(I)
     <,e1,e2,e3,e4,e5,e6,e7,e8,e9,e10

91    continue


931   DO 17 K=1,49
33    read(11,*)dayc(K),
c     <junk10,junk11,
     <N(K)

      DO 18 M=1,N(K)

15    read(11,*)junk,Ts(K,M)
c     <,junk12
18    continue

17    continue


      DO 41 K=1,48
      read(12,*)junk2,Rphin(K),junk3,junk4,junk5

41    continue


      ISSD(0)=0
      OSSD(0)=0
      QSSD(0)=0
      IISSD(0)=0
```

```
      SSSD(0)=0


      DO 888 K=1,48
      deldayc(K)=dayc(K+1)-dayc(K)


      TsKEL(K,1)=Ts(K,1)+273
      sIphia=0
      sIphib=0
      sphis=0
      sIphie=0
      sIphic=0
      sIphin=0



      sIIphia=0
      sIIphib=0
      sphis=0
      sIIphie=0
      sIIphic=0
      sIIphin=0



      sOphia=0
      sOphib=0
      sphis=0
      sOphie=0
      sOphic=0
      sOphin=0


      sQphia=0
      sQphib=0
      sphis=0
      sQphie=0
      sQphic=0
      sQphin=0






      sSphia=0
      sSphib=0
      sphis=0
      sSphie=0
      sSphic=0
      sSphin=0



c     Start of Imboden iterations
```

```
 DO 94 I=start1,start1+deldayc(K)-1


 dbtKEL(I)=dbt(I)+273

 es(I)=(2.171E8*(2.71**((-4157)/(Ts(K,1)+239.09))))*1.0002
 Iphib(I)=-(0.97*Isigma*(TsKEL(K,1)**4))
<*3.600*0.2388
 sIphib=sIphib+Iphib(I)

 Ophib(I)=-(0.96*Osigma*(TsKEL(K,1)**4))*2.388E-4
 sOphib=sOphib+Ophib(I)

 IIphib(I)=-(0.96*IIsigma*(TsKEL(K,1)**4))*0.2388
 sIIphib=sIIphib+IIphib(I)


 Qphib(I)=-(0.97*Qsigma*(TsKEL(K,1)**4))
 sQphib=sQphib+Qphib(I)



 Sphib(I)=-0.972*Ssigma*(TsKEL(K,1)**4)*0.8604
 sSphib=sSphib+Sphib(I)


 ea(I)=(2.171E8*(2.71**((-4157)/(dpt(I)+239.09))))*1.0002

 EL(I)=1.09*(1+(0.17*cloud(I)*cloud(I)))*1.24*(((ea(I))
</(dbtKEL(I)))**(0.14))
 Iphia(I)=(0.97*EL(I)*Isigma*(dbtKEL(I)**4))
<*0.2388*3.600
 sIphia=sIphia+Iphia(I)



 Ophia(I)=((Cat*Osigma*((dbtKEL(I))**6))*(1+(0.17*cloud(I)*cloud(I)
<))
<*0.97)*0.2388E-3
 sOphia=sOphia+Ophia(I)



 IIphia(I)=((1+(0.17*cloud(I)*cloud(I)))*0.937*IIsigma*0.97
<*(dbtKEL(I)**4))*0.2388
 sIIphia=sIIphia+IIphia(I)


 Qphia(I)=1.23E-16*(dbtKEL(I)**6)*(1+(0.17*cloud(I)*cloud(I)))*
<3600
 sQphia=sQphia+Qphia(I)



 Sea=2.1718E10*exp(-4157/((dpt(I)+273)-33.91))
 coef=1-cloud(I)
 IF(coef.le.0.4)then
 Sepsi=0.87-(coef*(0.175-(29.92E-6*Sea)))+2.693E-5*Sea
```

```
      ELSE
      Sepsi=0.84-(coef*(0.1-(9.973E-6*Sea)))+3.491E-5*Sea
      END IF
      Sphia(I)=0.97*Sepsi*Ssigma*((dbtKEL(I))**4)*0.8604
      sSphia=sSphia+Sphia(I)


      Iws(I)=(ws(I)*1000)/(60*60)
      f(I)=4.8+1.98*Iws(I)+0.28*(Ts(K,1)-dbt(I))
      Iphie(I)=(-f(I)*(es(I)-ea(I)))
     <*0.2388*3.600
      sIphie=sIphie+Iphie(I)


      Oes(I)=(es(I)*100*760)/(101.325)
      Oea(I)=(ea(I)*100*760)/(101.325)
      OE(I)=7.44E-5*ws(I)*(Oes(I)-Oea(I))
      Ophie(I)=(-998.2*(597.1-(0.57*Ts(K,1)))*((OE(I))/(1000*24)))
      sOphie=sOphie+Ophie(I)


      Sesw=2.1718E10*exp((-4157)/(TsKEL(K,1)-33.91))
      Sapress=apress(I)*100
      Seaa=2.1718E10*exp(-4157/((dbt(I)+273)-33.91))
      Tav=dbtKEL(I)/(1-(0.37*(Seaa/Sapress)))
      Twv=TsKEL(K,1)/(1-(0.37*(Sesw/Sapress)))
      IF(Twv.lt.Tav)then
      Sphie(I)=-(bstar*Iws(I))*(Sesw-Sea)
     <*0.8604
      ELSE
      Sphie(I)=-(Alambda*((Twv-Tav)**0.33)+bstar*Iws(I))*(Sesw-Sea)
     <*0.8604
      END IF
      sSphie=sSphie+Sphie(I)


      Qphie(I)=-(998.2*(597-(0.57*Ts(K,1)))*(a+(b*Iws(I)))
     <*((es(I)/1.0002)-(ea(I)/1.0002)))*3600
      sQphie=sQphie+Qphie(I)

c     Imberger et al does not specify,but it can be understood they
c     used CE-QUAL-R1 specifications for both IIphie and IIphic.

      IIphie(I)=Qphie(I)
      sIIphie=sIIphie+IIphie(I)




      Iphic(I)=(-0.642*f(I)*(Ts(K,1)-dbt(I)))
     <*0.2388*3.600
      sIphic=sIphic+Iphic(I)



      Sphic(I)=Sphie(I)*0.61E-3*Sapress*((Ts(K,1)-dbt(I))/(Sesw-Sea))
      sSphic=sSphic+Sphic(I)
```

```
      Ri(I)=-((9.81*(denair-denairTs)*z)/(denair*((Iws(I))**2)))

      IF(Ri(I))1111,1113,1112

1111  fRi(I)=(1-22*Ri(I))**(+0.80)

      go to 1120

1112  fRi(I)=((1+(34*(Ri(I))))**(-0.80))
      go to 1120

1113  fRi(I)=1
      go to 1120
1120  Ophic(I)=-fRi(I)*998.2*Cp*On*(apress(I)/P0)*Iws(I)*(Ts(K,1)-
     <dbt(I))*3600
      sOphic=sOphic+Ophic(I)

      Qphic(I)=-(998.2*(597-(0.57*Ts(K,1)))*(a+(b*Iws(I)))
     <*(CB+((apress(I)*1E-3)))*(Ts(K,1)-dbt(I)))*3600
      sQphic=sQphic+Qphic(I)

      IIphic(I)=Qphic(I)
      sIIphic=sIIphic+IIphic(I)

      Iphin(I)=Iphia(I)+Iphib(I)+phis(I)+Iphie(I)+Iphic(I)
      IIphin(I)=IIphia(I)+IIphib(I)+phis(I)+IIphie(I)+IIphic(I)
      Ophin(I)=Ophia(I)+Ophib(I)+phis(I)+Ophie(I)+Ophic(I)
      Qphin(I)=Qphia(I)+Qphib(I)+phis(I)+Qphie(I)+Qphic(I)
      Sphin(I)=Sphia(I)+Sphib(I)+phis(I)+Sphie(I)+Sphic(I)

      sIphin=sIphin+Iphin(I)
      sOphin=sOphin+Ophin(I)
      sQphin=sQphin+Qphin(I)
      sIIphin=sIIphin+IIphin(I)
      sSphin=sSphin+Sphin(I)
      sphis=sphis+phis(I)
```

```
91    continue
      AIphib=sIphib/deldayc(K)
      AIphia=sIphia/deldayc(K)
      AIphie=sIphie/deldayc(K)
      AIphic=sIphic/deldayc(K)
      Aphis=sphis/deldayc(K)
      AIphin=sIphin/deldayc(K)

      AIIphib=sIIphib/deldayc(K)
      AIIphia=sIIphia/deldayc(K)
      AIIphie=sIIphie/deldayc(K)
      AIIphic=sIIphic/deldayc(K)
      Aphis=sphis/deldayc(K)
      AIIphin=sIIphin/deldayc(K)


      AOphib=sOphib/deldayc(K)
      AOphia=sOphia/deldayc(K)
      AOphie=sOphie/deldayc(K)
      AOphic=sOphic/deldayc(K)
      Aphis=sphis/deldayc(K)
      AOphin=sOphin/deldayc(K)


      AQphib=sQphib/deldayc(K)
      AQphia=sQphia/deldayc(K)
      AQphie=sQphie/deldayc(K)
      AQphic=sQphic/deldayc(K)
      Aphis=sphis/deldayc(K)
      AQphin=sQphin/deldayc(K)




      ASphib=sSphib/deldayc(K)
      ASphia=sSphia/deldayc(K)
      ASphie=sSphie/deldayc(K)
      ASphic=sSphic/deldayc(K)
      Aphis=sphis/deldayc(K)
      ASphin=sSphin/deldayc(K)



      QDiff=ABS(Rphin(K)-AQphin)
      QSD=((QDiff)**2)
      QSSD(K)=QSD+QSSD(K-1)

      IDiff=ABS(Rphin(K)-AIphin)
      ISD=((IDiff)**2)
      ISSD(K)=ISD+ISSD(K-1)
```

```
      IIDiff=ABS(Rphin(K)-AIIphin)
      IISD=((IIDiff)**2)
      IISSD(K)=IISD+IISSD(K-1)

      ODiff=ABS(Rphin(K)-AOphin)
      OSD=((ODiff)**2)
      OSSD(K)=OSD+OSSD(K-1)



      SDiff=ABS(Rphin(K)-ASphin)
      SSD=((SDiff)**2)
      SSSD(K)=SSD+SSSD(K-1)



      WRITE(6,98)I,AIphia,AIphib,Aphis,AIphie,AIphic,
     <AIphin,
     <ISSD(K)
98    format(8(E12.5,3x))


      WRITE(7,198)I,AIIphia,AIIphib,Aphis,AIIphie,AIIphic,
     <AIIphin
     <,IISSD(K)

198   format(8(E12.5,3x))


      WRITE(8,298)I,AOphia,AOphib,Aphis,AOphie,AOphic,
     <AOphin
     <,OSSD(K)

298   format(8(E12.5,3x))



      WRITE(9,398)I,AQphia,AQphib,Aphis,AQphie,AQphic,
     <AQphin
     <,QSSD(K)

398   format(8(E12.5,3x))



      WRITE(14,1398)I,ASphia,ASphib,Aphis,ASphie,ASphic,
     <ASphin
     <,SSSD(K)

1398  format(8(E12.5,3x))




      start1=start1+deldayc(K)
888   continue
```

```
stop
end
```

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